Bulletin

of the

Geological Society of America

VOLUME 48 JULY 1, 1937 NUMBER 7

CONTENTS

PAGES

Regional contrasts in erosion in Indiana, with especial attention to the climatic factor in causation. By Stephen S. Visher.. 897–930

Thrusting in Huerfano Park, Colorado, and related problems of orogeny in the Sangre de Cristo Mountains. By W. S. Burbank and E. N. Goddard..... 931–976

Structure of the Honeybrook uplift, Pennsylvania. By George W. Stose 977–1000

Cambrian and Ordovician of northwestern Vermont. By Charles Schuchert 1001–1078

Subscription, $10 per year.

Publication Office: Florida Avenue and Eckington Place, Washington, D. C.

Communications for publication should be addressed to The Geological Society of America, Prof. Charles P. Berkey, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding part of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Washington, D. C., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PRESS OF JUDD & DETWEILER, INC., WASHINGTON, D. C.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 48, PP. 897-930, 27 FIGS. JULY 1, 1937

REGIONAL CONTRASTS IN EROSION IN INDIANA, WITH ESPECIAL ATTENTION TO THE CLIMATIC FACTOR IN CAUSATION

BY STEPHEN S. VISHER

CONTENTS

	Page
Introduction	898
Analysis of the erosion data	901
General statement	901
Wind erosion	902
Sheet erosion	903
Gullying	904
Combined sheet erosion and gullying	905
Summary	906
Some possible explanations of the regional erosional contrasts	907
Annual and seasonal rainfall	908
Glaciation and erosion	909
Relief and erosion	912
Climate and erosion	914
Regional contrasts	914
Temperature contrasts	915
Rainfall intensity contrasts	916
Average and excessive rainfall	916
Excessive short rains	918
Comparative erosion according to climatic regions	922
Lithology, structure, and erosion	924
Erosion and direction of slope	925
Summary	926
Conclusions	927

ILLUSTRATIONS

Figure		Page
1.	Distribution of soil erosion in Indiana	899
2.	Percentage of little soil erosion for each area	901
3.	Distribution of wind-erosion areas	901
4.	Distribution of moderate sheet erosion unaccompanied by gullying	902
5.	Distribution of considerable sheet erosion	902
6.	Distribution of severe sheet erosion	903
7.	Distribution of gullying	903
8.	Distribution of severe gullying	904
9.	Distribution of moderate or severe erosion by both sheet-wash and gullying	904
10.	Distribution of severe sheet erosion and severe gullying	908
11.	Distribution of precipitation from October to May	908
12.	Distribution of glaciation types	910

Figure Page
13. Relief map of Indiana ... 910
14. Relief map of Indiana ... 912
15. Distribution of freezing temperature ... 912
16. Distribution of snow-cover ... 916
17. Distribution of diurnal alternation of freeze and thaw ... 916
18. Distribution of one-day one-inch rainfall ... 917
19. Distribution of two-day two-inch rainfall ... 917
20. Distribution of two-day four-inch rainfall ... 918
21. Distribution of four-day six-inch rainfall ... 918
22. Distribution of ten-inch monthly rainfall ... 919
23. Distribution of excessive short rains ... 919
24. Distribution of April heavy rains ... 920
25. Distribution of May heavy rains ... 920
26. Distribution of greatest recorded six-day consecutive rainfall ... 922
27. Indiana climatic regions and subregions ... 922

INTRODUCTION

The Problem: How great are erosional contrasts in Indiana; what influences have produced them? The Method: An analysis of the soil erosion map recently issued by the United States Soil Conservation Service has been made; field observations have been made in all the counties of the State, in order to add supplementary data; maps showing the distribution of several factors presumably contributing to the causation of the contrasts were prepared and compared in detail with the erosion maps. Conclusions: Despite moderate differences with respect to elevation, relief, lithology, structure, geologic history, and climate, Indiana has notable regional contrasts in soil erosion. Factors causing these include contrasts in local relief, soil types, recency of glaciation, climate, and land use. Climatic contrasts clearly have played a larger part than has hitherto been appreciated. Regional contrasts in rainfall intensity—a factor hitherto almost ignored, because evidence as to its existence had not been compiled—appears to be of considerable significance in causing part of the observed erosional contrasts.

Regional contrasts in amount and type of erosion are of widespread geological interest, because such contrasts reflect the comparative importance or effectiveness of various geologic agencies.

Indiana, a representative mid-western State, has much greater contrasts in erosion than most geologists realize. The southern part of the State has much rather rugged land, and parts of the north, especially the northeast, are by no means level. Even in the generally level central and northwestern parts of the State, there has been considerable erosion locally.

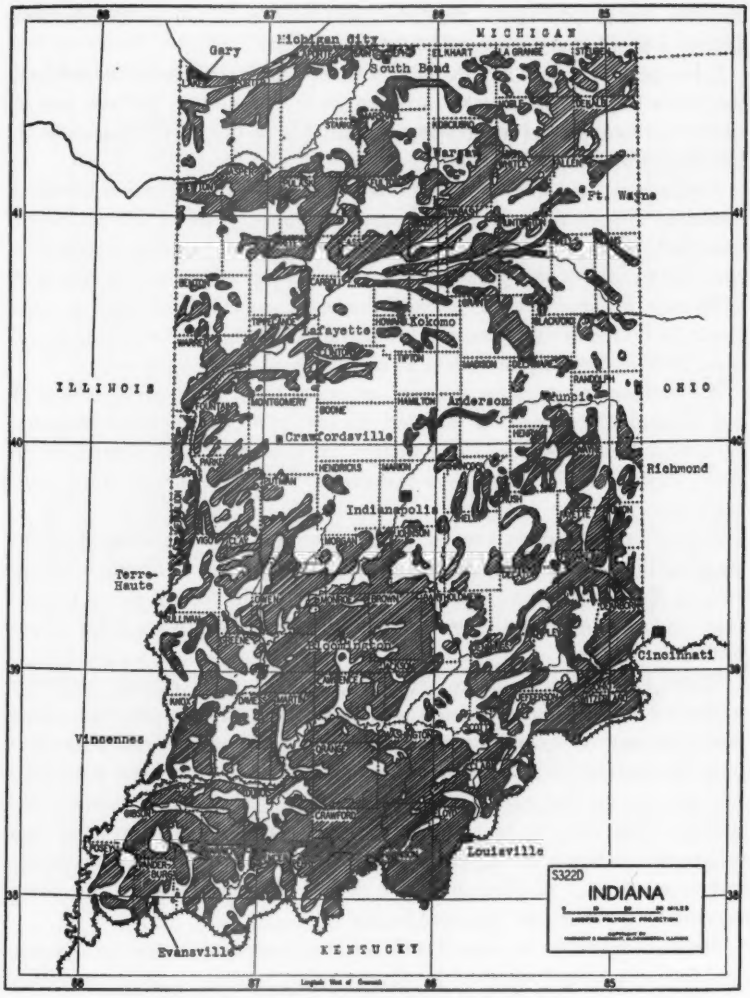


FIGURE 1.—*Distribution of soil erosion in Indiana*

Shaded areas are those in which considerable soil loss has been caused by erosion. Based on an erosion survey by the Soil Conservation Service.

In the following discussion, especial attention is given to soil erosion, rather than to erosion of all types, because a recent survey of the State mapped the amounts and the types of soil erosion. The coarser types of erosion are less accurately known, because little of the State has been

mapped topographically. However, soil erosion is, to a considerable degree, a preliminary phase of valley formation.

It has generally been assumed by those who have considered the subject, that such regional contrasts in erosion as are present in Indiana are of slight significance and are readily explainable as due to the contrasts in the glacial history of the State.

A summer's field work for the United States Soil Conservation Service,[1] including observations in all the counties, many years of miscellaneous observations in all parts of the State, and study of the map of soil erosion independently prepared by the Soil Conservation Service, made it evident to the writer that Indiana possesses considerable regional contrasts in both kind and amount of erosion, and that the explanation of these differences is by no means as simple as has been supposed.

As Indiana is, in many respects, representative of a large section of mid-latitudes, a somewhat detailed discussion of the regional erosional contrasts found there and the factors or influences that contribute to their explanation should have considerable utility, or at least suggestiveness.

About forty-three per cent of Indiana has undergone considerable erosion, having lost, on the average, more than one-fourth of its original topsoil, as disclosed by comparing the thickness of the topsoil found in borings in continuously well-forested tracts with the depth on tilled slopes and soils nearby.

Three types of erosion are extensive; namely, sheet erosion, gullying, and wind erosion. Sheet erosion is more widespread than gullying; about forty per cent of the State has undergone serious damage from sheet erosion, in contrast to about thirty-one per cent which has been somewhat gullied. As all the gullied areas are also subject to sheet erosion, the gullied thirty-one per cent of the State is included in the forty per cent where sheet erosion is considerable. Wind erosion is also of local significance, more than 650,000 acres, or nearly three per cent of the State, having been somewhat damaged by it.

From one-fourth to three-fourths of the original thickness of the topsoil has been widely removed by sheet erosion in almost exactly one-third of the State;[2] from an additional seven per cent of the State, more than three-fourths of the topsoil is generally gone. A fifth of the State has

---

[1] Other parts of this study were aided by grants from the Rollin D. Salisbury Memorial Fund of the University of Chicago and from the Research Fund of the Graduate School of Indiana University.

[2] A square mile is classified as of a given erosion type if it, "on the average," belongs to that erosion type. Thus, the statement does not mean that a large share of all the topsoil has been removed from every acre of one-third of the State, but, instead, that, in one-third of the area of the State, this type of erosion "prevails" (at least one-fourth of each square mile).

what the Soil Conservation Service officially calls "occasional" gullies, and one-ninth has "frequent" gullies, or severe gullying, and wind erosion has seriously damaged more than 30,000 acres.

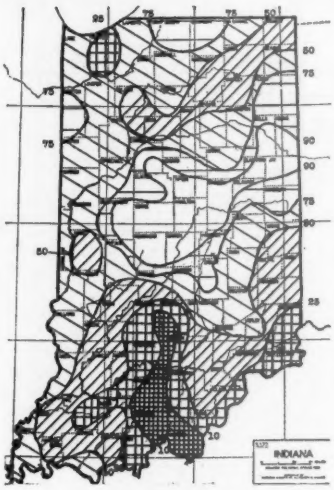


FIGURE 2.—*Percentage of little soil erosion for each area*

Figures indicate percentage of each county belonging to the little-erosion type. Based on a planimeter measurement of the erosion survey map.

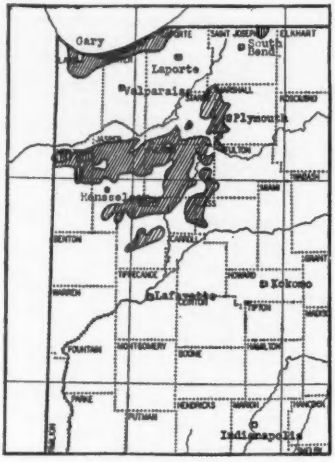


FIGURE 3.—*Distribution of wind-erosion areas*

Shaded areas indicate the larger tracts reported by the federal survey to have been notably damaged by wind erosion.

ANALYSIS OF THE EROSION DATA

GENERAL STATEMENT

The statements of the amount and kind of erosion here given are based on a study of the large-scale reconnaissance erosion-survey map already mentioned. Planimeter measurements of the areas of each type of erosion have permitted the calculation of the percentage of each erosion type in the total area of each county. These percentages are used as the basis for eight of the following maps and for the tables. The analysis of erosional contrasts may appropriately commence with the distribution and relative extent of the areas upon which, in general, the soil erosion has been negligible. Figure 1 shows the distribution of two types of erosion. The unshaded areas have little erosion as compared with the shaded areas, which have considerable erosion. Within the shaded and unshaded areas, there are small areas of the opposite type, too small to be shown on this scale. This map indicates that most of the extensively eroded land is in the southwestern third of the State, in parts of the southeast,

and in various wide bands in the north. Conversely, the areas that are little eroded are largely in the central part of the State.

Figure 2 shows the percentage of each area, based on data for each county, classed as subject to negligible erosion. This map indicates some

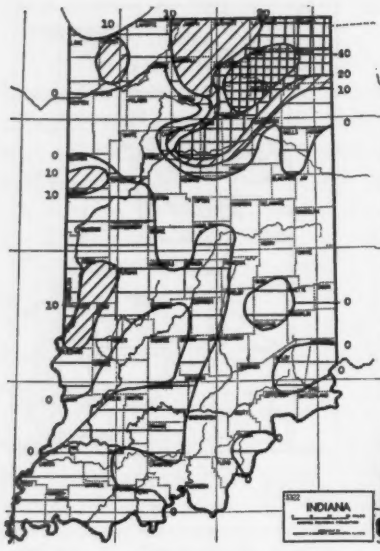


FIGURE 4.—*Distribution of moderate sheet-erosion unaccompanied by gullying*

Moderate sheet-erosion designates those areas where one quarter, or more, of the topsoil is gone. Figures indicate percentage of the area which belongs to this type. Figures based on planimeter data by counties.

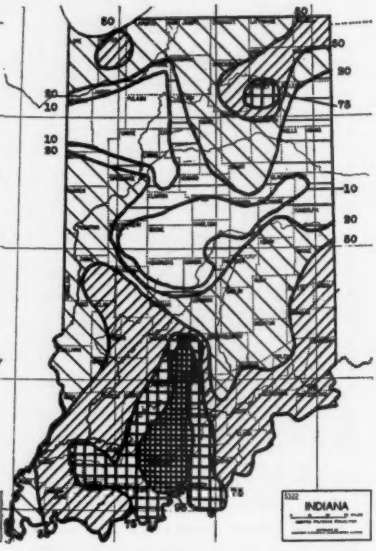


FIGURE 5.—*Distribution of considerable sheet-erosion*

Figures indicate the percentage of the area having either moderate or severe sheet-erosion. Figures based on planimeter measurements of the erosion map.

of the sharp regional contrasts more effectively than does Figure 1. The chief areas in which less than one quarter of the land is classed as being notably eroded are in the central and northeast center, the central northern, and a small area in the northwest part of the State.

WIND EROSION

Figure 3 shows the areas which have been damaged appreciably from wind erosion. Although some small sandy areas elsewhere, especially in Jackson County and in the southwest (Knox and Gibson counties), have lost some topsoil from wind erosion, the only counties that have had as much as one per cent of their area thus damaged are near the northwestern corner of the State.

SHEET EROSION

Figure 4 treats of sheet erosion, unaccompanied by any appreciable amount of gullying. In the northeastern counties, from ten to forty per cent of the area has generally lost one-fourth or more of its original top-

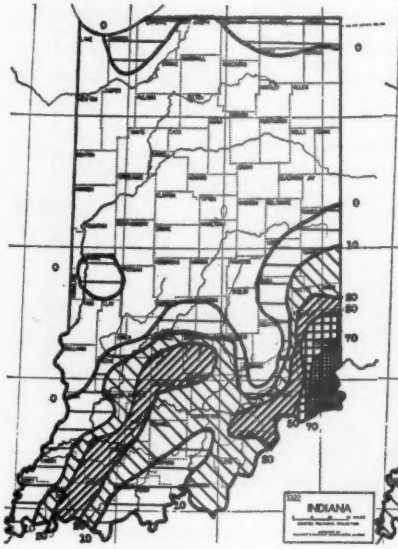


FIGURE 6.—*Distribution of severe sheet-erosion*

Severe sheet-erosion designates those areas where three quarters, or more, of the topsoil is gone. Figures indicate the percentage of the area in which this type prevails.

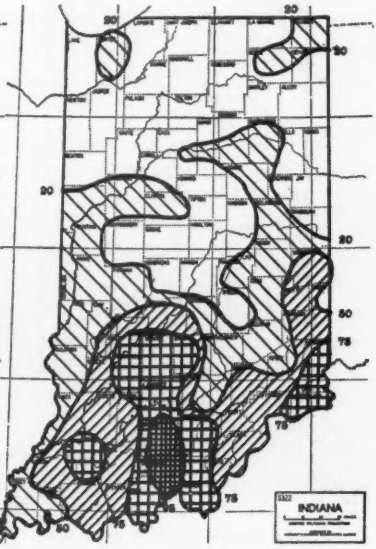


FIGURE 7.—*Distribution of gullying*

Including moderate and severe gullying. Figures indicate the percentage of the area belonging to the type "with occasional gullies" and to the type "with frequent gullies."

soil by sheet erosion without gullying. Except in two small areas—one in the northwest and the other in the west-central part of the State—no other counties have simple sheet erosion upon as much as ten per cent of their areas; indeed, most counties have less than one per cent of their land eroded in that way.

Figure 5 shows sheet erosion, with and without gullying. It is very different from Figure 4, because most sheet erosion is accompanied by gullying. Half, or more, of the area of each of the southern counties, and 75 per cent, or more, of several south-central counties have undergone considerable (about fifty per cent of the topsoil gone) sheet erosion.

Figure 6, which illustrates severe sheet erosion (three-fourths, or more, of the original topsoil gone, on the average), reveals that this type of ero-

sion is confined to the southeastern third of the State, except for four counties on the Michigan border. It is worst at the southeast corner, but farther west are five counties in which it is almost as bad.

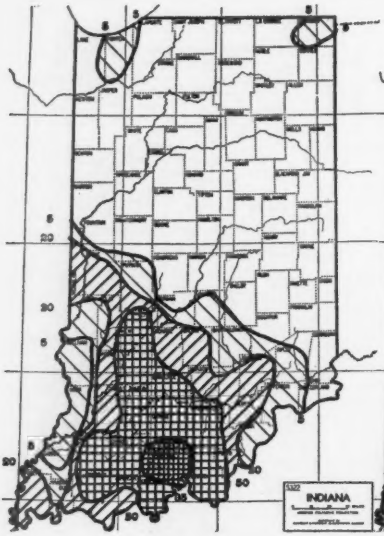


FIGURE 8.—*Distribution of severe gullying*

Figures indicate percentage of area classed by the erosion survey as generally having numerous gullies.

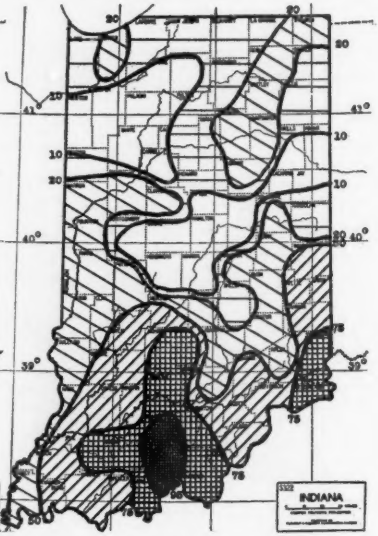


FIGURE 9.—*Distribution of moderate or severe erosion by both sheet-wash and gullying*

Figures indicate percentage of area in which these types prevail.

GULLYING

Figure 7 shows the distribution of gullies. Gullying is widespread; only four counties were mapped by the erosion survey as lacking sizable gullied areas. Three additional counties have, however, only about one per cent of their land somewhat gullied, and twelve other counties have less than a tenth of their area of this type. By contrast, most of the counties of the southern half of the State have forty per cent, or more, of their land somewhat gullied; indeed, a dozen counties—nearly all in the central part of southern Indiana—have three-fourths of their land classified as somewhat gullied.

Two types of gullying were distinguished by the erosion survey; namely, "occasional" and "frequent" gullies. The percentage of the land with "occasional" gullies is greatest at the extreme southeast, where nine counties have more than half of their area in this type. These extend from

Wayne to Clark counties. Switzerland, Ohio, and Dearborn counties have 77 to 79 per cent of their land of this type.

Figure 8 shows the percentage of the area classed as having "frequent" gullies. This map reveals that severe gullying is largely confined to the southwestern third of the State. Ten counties near the center of that section have more than half of their land of this type. On the other hand, more than half of the counties of the State have no considerable tracts of badly gullied land, and fifteen other counties have less than five per cent of their land thus classified.

COMBINED SHEET EROSION AND GULLYING

Gullying and sheet erosion generally occur simultaneously on the same land. Figure 9 shows the percentage of each area having, in general, lost more than one-fourth of the original topsoil and having also numerous gullies. This map shows that more than half of the area of most of the southern counties was classed by the federal survey as belonging to this type. Some ten south-central and three southeastern counties have more than 75 per cent of their areas in this category, and, near the center of the south-central area, Crawford and Orange counties have 99 per cent and 95 per cent, respectively. This does not mean that 99 acres out of every 100 in Crawford County are this badly eroded, but that almost the entire county belongs to the type in which such severe erosion prevails.

Maps of the various combinations of sheet erosion and gullying have been made, but there is not space here to reproduce all of them. Three may, however, be briefly described. The counties with the largest percentage of their area damaged by moderate sheet erosion and occasional gullies, and the percentage of their area thus affected, are Brown 66, Orange 65, Clark 55, Morgan 54, Gibson 50, Jefferson 49, Knox 45, Henry 41, Warren 40, Fayette and Wayne 39, Clay 35, Decatur and Monroe 34, Sullivan 33, Union and Tippecanoe 32, Grant 31, and Jennings and Delaware 28. Most of these counties are in the southwestern third of the State. The largest area with little erosion of this particular type is in the northwestern part of the State; the next largest is in the center.

Moderate sheet erosion is accompanied by numerous gullies in much of the southwestern part of the State, where twelve counties, extending from Parke to Floyd and Warrick, have thirty per cent, or more, of their areas of this type, and four counties have more than half of their areas thus classified. By contrast, only three counties to the north or east of a line from Jefferson to Fountain counties have as much as one per cent of their land in this class. They are Porter (40), Steuben (7), Cass (5), and Huntington (3).

Severe sheet erosion and "occasional" gullies damage most land at the extreme southeast, where four counties have more than half of their

land in this category; namely, Ohio 76, Dearborn 73, Switzerland 69, and Franklin 58. Nearby counties also ranking relatively high are Fayette 21, Union 10, Wayne 11, and Ripley 4. In no other part of the State is this combination extensive except in Vanderburg County (21 per cent), Lawrence (5 per cent), Harrison (4 per cent), and Washington (3 per cent). None of the counties of the northern half of the State has as much as one per cent of its land of this type.

The final map of erosion is that showing the most serious combination, that of severe gullying and severe sheet erosion. In this classification, less than one-fourth of the original thickness of the topsoil remains, on the average, and in many places the topsoil is all gone. Likewise, gullies are numerous, a considerable number of them cutting well into the subsoil, or even down to bed-rock. This most extreme type of widespread erosion in Indiana (Fig. 10) is largely confined to the southern third of the State and affects the largest percentage of the land in the counties extending from Brown to Pike, which have from 20 per cent to 52 per cent of their land of this category. Various nearby counties also rank high. North of the line marked 0 in Figure 10, there are, however, only nine counties with appreciable erosion of this type. They, and the percentages of their areas, thus affected, are Wayne 5, Steuben and Laporte 3, Porter and Lagrange 2, Noble and Henry one, and Parke and Elkhart with less than one.

SUMMARY

Except for wind erosion, which is most extensive in the northwest, erosion is most extensive in the southern part of the State. Indeed, much of the central and north-central region has little serious erosion. In the extreme northern and northeastern counties, moderate sheet erosion without gullying is, to be sure, rather widespread in the rougher areas. Nevertheless, sheet erosion in general is most extensive in the counties near the Ohio River. Gullying similarly increases in severity southward, although it is by no means lacking on the steeper slopes in the north. Gullies are most numerous, on the average, in the south-central and southwestern parts of the State, although, on some of the almost flat land near the rivers, they are not numerous.

The combination of considerable (moderate or severe) sheet erosion with considerable gullying (occasional or frequent gullies), shown in Figure 9, indicates clearly the general southward increase shown by most of the maps. This combination is of such broad significance that it has been analyzed in a special way. The 17 counties north of latitude 41° north have, on the average, about one-eighth of their areas classed as subject to considerable erosion of these types; the belt between the 40th and 41st parallels (22 counties) has an average of about one-sixth of its area

of these types; the 25 counties between the 39th and the 40th parallels have an average of 43 per cent of their areas so classed. The 24 counties south of the 39th parallel, have, on the other hand, an average of 66 per cent of their areas in this classification.

SOME POSSIBLE EXPLANATIONS OF THE REGIONAL EROSIONAL CONTRASTS

Why this southward increase? Why greater erosion in the central section of southern Indiana than on either side? Why the intense erosion at the extreme southeast corner? Why more gullying at the southwest than in the southeast or even in the south-central section?

The regional contrasts in erosion must necessarily correlate with variations in the effectiveness of the agencies of erosion, which here are wind and running water, and with the availability of materials that can be carried away by these agents. The effectiveness of the wind varies with its strength (velocity and gustiness), with the dryness and looseness of the soil, and with the protection afforded by vegetation or other barriers to wind movement. In Indiana, wind erosion is largely confined to sandy soils, which are the only ones that normally are loose when dry, loams and clays commonly being quite hard when dry. Even when pulverized by cultivation, clays and loams in Indiana seldom remain loose on top for more than a short time. The reason is that showers are frequent enough, especially in spring when most cultivation is done, so that a crust is quickly formed on top of the man-loosened layer of clay or loam. Sand, however, lacks the stickiness required to form such a crust; hence, soon after a shower, it may be loose enough to be moved by the wind. However, sandy soils in all parts of the State are not equally eroded by the wind. Sandy soils in southern Indiana are eroded less than are similar soils in the northwest, partly because they are less dry in winter. The northwestern corner of the State receives less precipitation in winter than any other part of the State, only about half as much as southern Indiana. Moreover, it also has the strongest winter winds. These conditions are highly favorable to wind erosion, as are the extensive sandy areas, the small interference from trees or hills, and the extensive corn fields, many of which are barren in winter.

Most erosion in Indiana is accomplished by the run-off of rain. The greater the run-off, the greater the potential erosion, both because of the greater volume of water to do the work, and, especially, because of the greatly increased velocity which the run-off attains as the amount increases. Where there is little run-off, the friction with the surface and with small obstacles normally permits only a slow flow, which means little transporting power and, hence, little erosion.

The amount of run-off varies with six main factors: (1) amount of rainfall, (2) its rate or intensity, (3) the slope or degree of roughness, (4) amount of absorption by the soil and associated materials, (5) evaporation, and (6) vegetal cover. The slope includes not only the larger features, but also the lesser irregularities and the presence of channels. The amount of absorption depends upon the porosity of the soil itself, upon the openings in the soil, such as root cavities and the burrows of animals, and especially upon the degree of saturation. Even a sand, which can absorb one-fourth of its weight of water when dry, cannot readily absorb much water if its upper layers are already saturated.

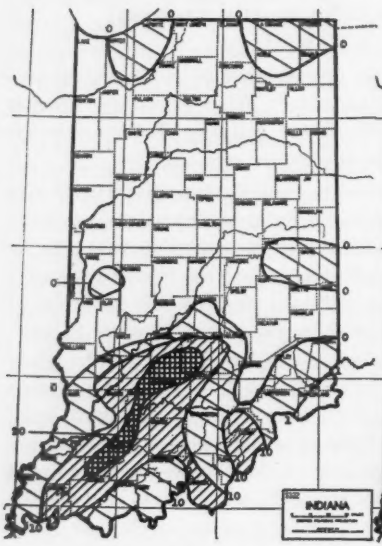


FIGURE 10.—*Distribution of severe sheet-erosion and severe gullying*

Figures indicate percentage of area in which these types prevail.

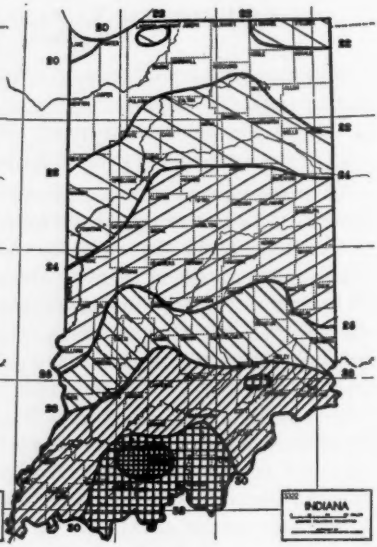


FIGURE 11.—*Distribution of precipitation from October to May*

Figures show total inches, based on all available data from 86 Indiana stations, covering the period from 1898 to 1932.

ANNUAL AND SEASONAL RAINFALL

In Indiana the annual average rainfall varies from slightly less than 35 inches, at the north, to about 45 inches, at the south. Hence, if all other factors were equal, the amount of erosion should be approximately one-third greater at the south than in the north. But, instead of being one-third greater, it is several times as great.

Rainfall during the four warmer months, from June to September inclusive, causes much less erosion, on the average, than do similar amounts during the cooler months, falling at similar rates. The reason for this is that in summer the soil is commonly dry enough to absorb a considerable share of all the rainfall, if it does not fall with too great intensity. In summer, evaporation is relatively rapid, soon removing part of the water. Furthermore, the soil is more porous in summer than during the rest of the year, because of the greater activity of burrowing animals and man (cultivation). Finally, and of especial importance, vegetal protection normally is greatest in the warmer months.

Hence, the regional contrast in the amount of rainfall during the cooler months is far more significant, so far as erosion is concerned, than is the annual total. All parts of Indiana receive about the same amounts of rainfall during the summer; hence, the notable annual contrast between northern and southern Indiana is largely due to cool season-contrasts. During the eight cooler months the precipitation totals in the north average about 21 inches, in the center about 25 inches, and at the south about 30 inches. Figure 11 shows the "standard" precipitation [3] totals for these months for the various parts of the State.

This map shows that southern Indiana receives nearly a half more precipitation during the cooler eight months than does the north. Indeed, several stations in south-central Indiana receive approximately 32 inches of rain in these months, which is 60 per cent more than the totals for three northwestern and two northeastern stations.[4]

During the three winter months, the contrast is even greater, because of the scanty precipitation received then at the north. In winter the southern counties receive nearly twice as much precipitation as does the north, and Crawford County and vicinity more than twice as much as the northwestern or northeastern corners of the State.

Hence, regional contrasts in the amount of rainfall received annually, and especially that received during the cooler months, help to explain regional contrasts in erosion. But erosional contrasts are much greater than contrasts in mere annual or seasonal rainfall; hence, other factors merit consideration.

GLACIATION AND EROSION

Southern Indiana differs from northern Indiana with respect to glaciation. Figure 12 is a glacial map, simplified from Malott's 1922 map,

[3] The "standard data" are the figures for the 35-year-period, 1898-1932, and are more strictly comparable than are the averages of all data.

[4] Snowfall to the equivalent of three or four inches of rain is received each winter in northern Indiana; southern Indiana receives about half as much, or less. Hence, the difference in the amount of winter rain is somewhat greater than is suggested by Figure 11 and the associated discussion, which is concerned with all precipitation.

which, in turn, was based on Leverett's 1914 map.[5] When this map is compared with the erosion maps, it is apparent that there are general correlations between glaciation and erosion. Thus, the unglaciated area [6] has, in general, more erosion than has the glaciated, and the areas cov-

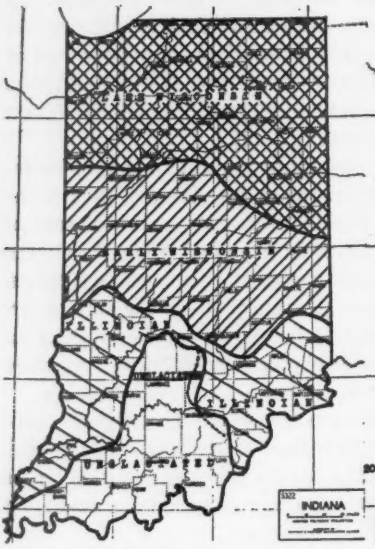


FIGURE 12.—*Distribution of glaciation types*

In the area labelled unglaciated, that part east of the dashed line is not entirely lacking in glacial deposits, although most of it is rugged upland, resembling the unglaciated area to the west rather than the more strongly glaciated area to the east.

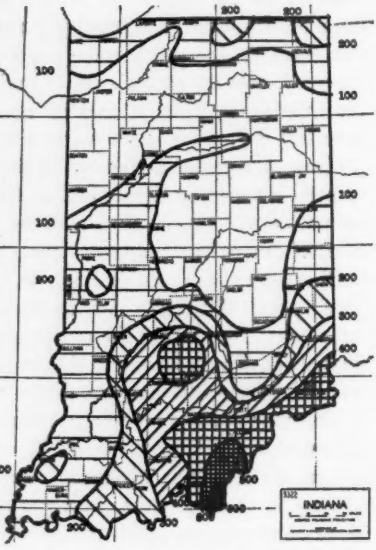


FIGURE 13.—*Relief map of Indiana*

Based on maximum local differences in elevation for each county. Figures indicate feet of difference.

ered by the older, Illinoian, glacier have more erosion than has the section covered by the Wisconsin ice sheet. Similarly, the area covered by the Late Wisconsin advance has, on the average, less water erosion than does the Early Wisconsin drift sheet. Such contrasts are to be expected, as erosion requires time, and sufficient time has not yet elapsed since the Late Wisconsin to drain adequately many parts of northern Indiana. Run-off erosion is necessarily slow where there are no appreciable slopes or channels. On the other hand, the region covered only by the Illinois

[5] Frank Leverett: *Glacial map of Indiana*, U. S. Geol. Surv. (1914).
C. A. Malott: *Physiography of Indiana*, pt. 2, *Handbook of Indiana Geology*, Ind. St. Geol. Surv. (1922) pl. 3.

[6] W. D. Thornbury: *Notes on the glacial boundary in southern Indiana*, Ind. Acad. Sci., Pr., vol. 41 (1932) p. 351-354.

ice sheet has been fairly well drained, and has been reduced to slopes. Most of the unglaciated area is hilly.

The extent of the correlation between glaciation and erosion is suggested by the data summed up in Table 1, which is a study of the percentages of the area of each glacial type that belong to the various erosion types. It reveals that the part of the State covered by the Wisconsin drift sheets is less eroded than is the part that was glaciated by the Illinoian glacier, with respect to eight of the nine types or combinations of erosion analyzed. Likewise, the unglaciated area is more eroded than is the glaciated area on the average. It shows, however, that in four of the nine types of erosion analyzed, the Late Wisconsin area is more eroded than is the early Wisconsin, and in respect to two types, the Illinoian area is more eroded than the unglaciated area.

TABLE 1.—*Soil erosion correlated with glaciation types*

Erosion type	Percentage			
	Late Wisconsin	Early Wisconsin	Illinoian	Unglaciated
Little erosion	63.0	71.0	42.0	28.0
Sheet erosion without gullies	13.7	3.4	1.8	2.2
Moderate sheet erosion with or without gullies	28.0	26.0	36.0	56.0
Considerable sheet erosion (moderate or severe)	28.0	28.0	57.0	71.0
Severe sheet erosion	0.4	2.3	21.3	15.2
Considerable gullying (moderate or severe)	14.0	25.0	56.5	70.0
Severe gullying	2.4	4.0	15.4	46.5
Considerable sheet erosion and gullying (moderate or severe of both)	14.4	25.4	56.0	73.6
Severe sheet erosion and severe gullying	0.4	0.6	5.3	13.2

Although both the southeastern and the southwestern parts of the State were covered by the Illinoian ice sheet, they differ appreciably in erosion type; the percentage of the area with severe sheet erosion is much greater in the east, and gullying is much more widespread and severe in the west. Similarly, within each of the other glacial types there are substantial differences in erosion. Moreover, parts of the unglaciated area are little eroded, and parts of the most extensively glaciated area, the Late Wisconsin, are badly eroded.

Thus, although it is evident that glaciation has been an important factor in the formation of erosional contrasts in Indiana, glacial contrasts are obviously inadequate to explain fully the erosion contrasts.

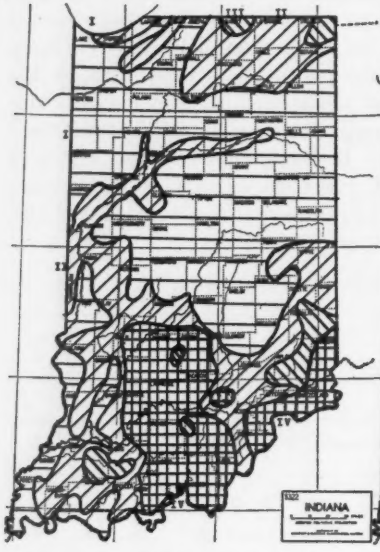


FIGURE 14.—*Relief map of Indiana*

Showing relative ruggedness. (I) indicates fairly level areas (local relief generally less than 50 feet); (IV) indicates rugged areas (local relief often more than 200 feet); (II) and (III) indicate intermediate areas.

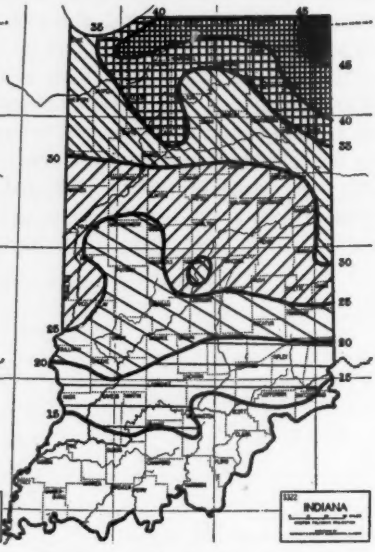


FIGURE 15.—*Distribution of freezing temperature*

Figures indicate days per average year when there has been freezing all day. Based on records of 52 stations for 13 years.

RELIEF AND EROSION

Another factor of unquestionably major significance is the amount of slope of the land. Figures 13 and 14 show local relief. Figure 13 is based on the maximum local differences in elevation for each county; Figure 14 divides the State into four categories of ruggedness. The data upon which both maps were based are rather unsatisfactory, as topographic maps of only a small portion of Indiana have been made. Figure 13 is based on data compiled by Malott.[7] Figure 14 is based on miscellaneous field impressions. The part of the State shown as cross-shaded is, in general, rather hilly; the area which is shaded most lightly is, in general, level or almost level. The other two types are intermediate, the darker ones being the rougher.

Some of the erosional contrasts clearly are partly due to local relief,

[7] C. A. Malott: *op. cit.*

as, for example, is part of the considerable contrast between the southeastern and the southwestern corners of the State, both formerly covered by the Illinoian ice sheet, but now differing markedly in local relief. The considerable local relief also helps to explain the presence of extensive erosion in the area from Parke County to Owen County, although it was covered by the Illinoian ice sheet. Likewise, the greater erosion in some of the northern counties, especially at the northeast, as compared with adjacent counties to the south, is obviously related to greater relief, due to terminal moraines.

The correlation between relief and extent of erosion is suggested by Tables 2 and 3, based, as was Table 1, on the percentage of each of the various erosion types in the counties that are entirely or chiefly in the various relief types. Table 2 deals with three general regions so far as

TABLE 2.—*Soil erosion correlated with topography*

Erosion type	Percentage		
	Regions prevailingly level to gently rolling	Areas prevailingly rolling with some hilly areas	Regions prevailingly hilly
Little erosion	70.0	50.5	20.0
Moderate sheet erosion without gullying	7.3	7.9	0.7
Moderate sheet erosion (with or without gullies)	24.2	38.7	54.6
Considerable sheet erosion (moderate or severe)	24.7	50.6	76.0
Severe sheet erosion	0.6	11.7	21.7
Considerable gullying (moderate or severe)	18.0	42.8	72.0
Severe gullying	5.4	15.7	38.8
Considerable sheet erosion and gullying	18.4	42.7	75.3
Severe sheet erosion and severe gullying	0.4	6.6	8.6

ruggedness is concerned. It shows that the type that is most nearly level has the largest percentage of land that is little eroded, and has less sheet erosion and less gullying than the more rugged types. However, the three relief types do not differ in erosion to the extent that would be true if relief were the dominant factor, as is shown, for example, by the fact that, in respect to several types, there is little difference in the amount of erosion in different relief regions.

Table 3 concerns relief contrasts based on the maximum local relief in each county. This table reveals that the greater the relief, the smaller the percentage of the land that is subject to little erosion of most types. But, again, there are exceptions, which prove the inadequacy of relief contrasts to explain the erosion contrasts.

TABLE 3.—*Soil erosion correlated with maximum local relief*

Erosion type	Percentage of local relief			
	Less than 100 feet maximum	From 100 to 200 feet	From 200 to 400 feet	From 400 to 600 feet
Little erosion	70.4	56.0	36.5	25.2
Sheet erosion without gullies	9.0	7.2	2.7	0.1
Moderate sheet erosion with or without gullies	21.6	35.2	48.9	42.5
Considerable sheet erosion (moderate or severe)	22.8	40.0	64.6	72.0
Severe sheet erosion	0.1	6.0	14.9	29.5
Considerable gullying (moderate or severe)	14.0	34.1	34.9	71.4
Occasional gullies	12.6	21.8	29.6	42.5
Considerable sheet erosion and gullying	13.1	33.3	61.9	71.6
Severe sheet erosion and severe gullying	0.0	3.5	7.3	10.3

CLIMATE AND EROSION

REGIONAL CONTRASTS

Regional contrasts in annual and seasonal totals of rainfall have already been mentioned as factors contributing to regional contrasts in the work of both wind and running water. There are, however, many additional regional contrasts in climate that influence erosion. Some of these are: (1) Contrasts in the duration of frozen ground—that is, the percentage of the year when the ground is continuously frozen and, hence, rather well protected from erosion; (2) contrasts in the duration of snow-cover; (3) contrasts in the frequency of diurnal alternation of freezing and thawing. Repeated freezings, and of even more importance, repeated thawings, are conducive to soil movement both by slumping and by run-off.

Regional contrasts in the intensity of rainfall are of especial significance to erosion and that data has been analyzed to show (4) the frequency of daily rains of one inch or more; (5) daily rains of two inches or more; (6) two-day rains of two inches or more; (7) six inches or more of rain falling within four days or less; (8) "excessive rains of short duration" (less than 18 hours); (9) as soil erosion is especially rapid whenever heavy rains occur while the soil is still loose after plowing, especial attention has been given to the frequency of excessive rains during the spring months. Finally, the greatest amounts of rainfall ever recorded in six consecutive days are considered.

TEMPERATURE CONTRASTS

Figure 15, the duration of freezing temperature, shows that in northern Indiana during about forty days per average year the ground is continuously frozen and, hence, not subject to run-off erosion. On the other hand, in central Indiana the soil is thus protected during an average of about 30 days; in southern Indiana, official, shade, temperatures are continuously below freezing on only about 15 days, or only about a third as long as at the north. Hence, the regional contrasts in this climatic factor help to explain the greater erosion at the south.

Figure 16, the average duration of snow-cover, is based on rather unsatisfactory data, as the snow-cover soon becomes patchy wherever the fall is as slight as it normally is in southern Indiana. Parts of northern Indiana receive, however, an average of more than forty inches of snowfall, or three times as much as is usual in the southern part of the State. The differences are even greater upon south-facing slopes. The thin snow-cover characteristic of southern Indiana melts much sooner on such slopes than does the deeper snow of northern Illinois.

Figure 17, the frequency of diurnal alternation of freezing and thawing, is of considerable interest, as repeated thawing induces solifluction or creep on sloping fields, and greatly increases subsequent run-off erosion, because surface freezing greatly loosens the soil, and also breaks some of the roots of shallow-rooted plants, such as winter wheat. Figure 17 is based on the differences between the annual number of days with temperatures continuously below freezing (Fig. 15) and the number of days with frost (not here published; about 125 days a year have frost in most of the northern half of the State, in contrast to about 90 days at the south). The regional contrasts revealed by Figure 17 are small but are slightly in favor of southern Indiana as compared with central Indiana. Northern Indiana, however, is favored as compared with central Indiana. Field observations reveal, however, that, in southern Indiana, south-

facing slopes have many more alternations of freeze and thaw than Figure 17 indicates and notably more than do similar slopes in northern Indiana. This is because (1) southern Indiana is near the margin of the region of continuously freezing temperatures; (2) the ground seldom has more than

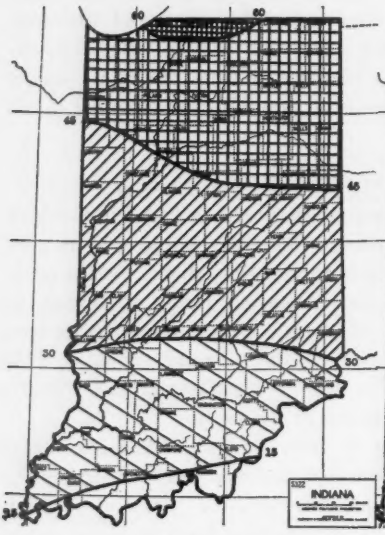


FIGURE 16.—*Distribution of snow-cover*

Figures indicate days per average year. Based on all available records at 5 regular full time, weather bureau stations in Indiana, supplemented by corresponding data gathered at the regular stations at Chicago and Peoria, Illinois; Louisville, Kentucky; Cincinnati and Dayton, Ohio.

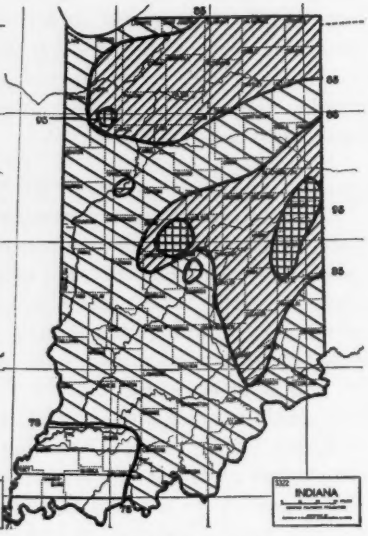


FIGURE 17.—*Distribution of diurnal alternation of freeze and thaw*

Figures indicate days per average year having temperature fluctuations passing the freezing point. Based on all data for the decade 1926-1935 for 22 well-distributed stations, mostly rural.

a thin mantle of snow, which is soon melted from the south slopes; and (3) insolation is appreciably more effective than it is farther north during the months when the sun is relatively low in the sky at noon, and nights are long enough to facilitate nocturnal freezing.

RAINFALL INTENSITY CONTRASTS

Average and Excessive Rainfall.—Figure 18, the frequency of daily precipitation of one inch or more, shows that, during the winter months, such rains occur in the southwestern part of the State three times per average winter, which is more than three times as frequently as in northern Indiana. Indeed, the northeastern section receives such rains less than one-sixth as often as does southwestern Indiana.

Figure 19 shows the distribution of areas with a frequency of two inches or more of rain falling within two consecutive days in winter. A large share of such rainfall must necessarily run off, because the ground normally is saturated by the first inch, or less, of winter rain. This map

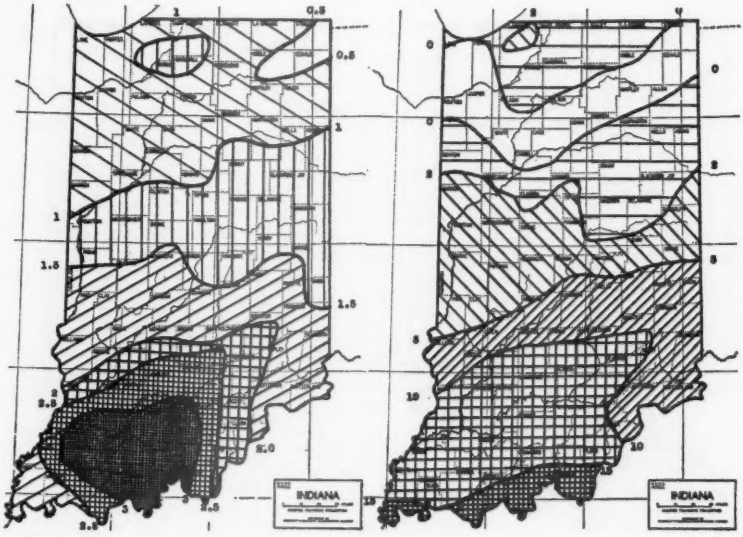


FIGURE 18.—*Distribution of one-day one-inch rainfall*

Figures indicate times per average winter. Based on daily records from 1926 to 1935, covering 85 stations.

FIGURE 19.—*Distribution of two-day two-inch rainfall*

Figures indicate times per average winter. Based on daily records from 1926 to 1935, covering 85 stations.

shows that such rains are from several times to many times as frequent in southern Indiana in winter as in most of northern Indiana. Indeed, in a considerable part of the north such rains occur less than once a decade, instead of once each winter as in the southwest.

Rainfalls of much more than two inches in two consecutive days are infrequent in Indiana, but the regional variations in their frequency are striking. For example, of four inches of rain, or more, falling within two consecutive days, the southwestern corner receives one such downpour each three years on the average, whereas, in the northeast, such rains occur less than one-sixth as often (Fig. 20).

Figure 21 shows the distribution of even greater totals of rain, but for a longer period; namely, six inches, or more, falling within four consecutive days. Such prolonged heavy rains necessarily result in much run-off and, consequently, in much erosion if other conditions are favorable. This

map shows that, in south-central Indiana, six-inch rains in four days, or less, occur in from 10 to 15 per cent of the years, whereas a large area in northern Indiana has not experienced such rains since official records have been kept—forty to fifty years for most stations.

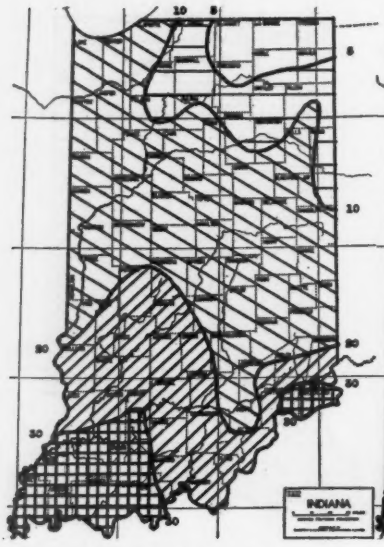


FIGURE 20.—*Distribution of two-day four-inch rainfall*

Figures indicate percentage of years having such rains. Based on all available data (from the beginning of official records up to 1937, an average of about 45 years of records, but some records are for more than 60 years) from 86 stations.

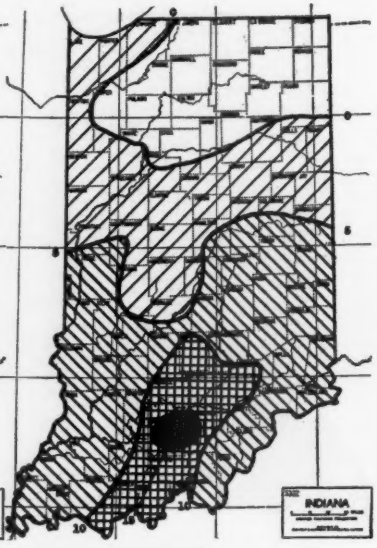


FIGURE 21.—*Distribution of four-day six-inch rainfall*

Figures indicate percentage of years having such rains. Based on all available data from 86 stations, for the duration of their official records, an average of about 45 years.

Data as to a still more prolonged type of heavy rainfall—namely, the frequency of monthly totals of large amounts—are readily available in the weather bureau records. Figure 22 shows that monthly totals of ten inches, or more, during the cooler half year occur at least once a decade in the southwest, but almost never in the northern part of the State. The area where erosion is most intense (Fig. 10) roughly coincides with the area that has most frequent monthly rainfall totals of ten inches, or more.

Excessive Short Rains.—Consideration has now been given to annual, seasonal, and monthly rainfalls, and to the rainfalls during one day, two days, and four days. Another type of rainfall that is highly significant so far as soil erosion is concerned is the "short rains of excessive amounts."

When an inch of rain falls within one hour, or an inch and a half within two hours, for example, the rainfall is classed as "excessive," and the run-off normally is great from slopes, as there is not time for much rain to soak into the ground. Such rains during the cooler season are especially

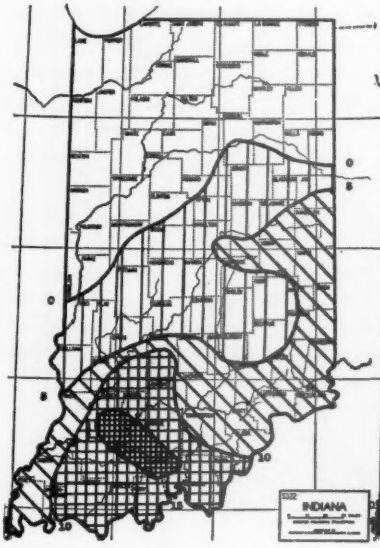


FIGURE 22.—*Distribution of ten-inch monthly rainfall*

Figures indicate percentage of years with cooler half years having such rains. Based on all available data from 86 stations. (Duration of record averages about 45 years.)

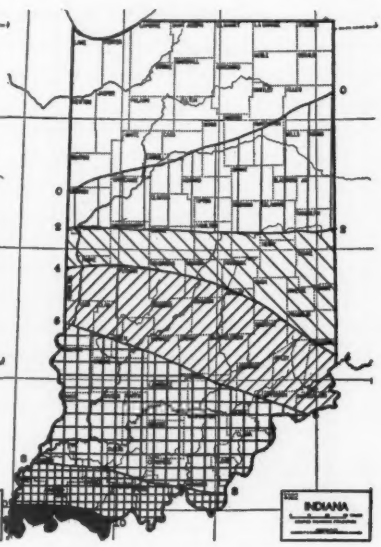


FIGURE 23.—*Distribution of excessive short rains*

Excessive short rains designates rains of one inch per hour, one and a half inches per two hours, or more. Figures indicate times per thirty winters (1903 to 1932). Based on all available records at 5 regular weather bureau stations in Indiana, supplemented by data gathered at regular stations at Chicago and Peoria, Illinois; Louisville, Kentucky; Cincinnati and Dayton, Ohio.

conducive to soil erosion, because the soil normally is already fairly well saturated before the deluge commences. Figure 23 shows that excessive short rains are almost unknown in winter in northern Indiana, but occur once in about every four winters in the southwest.

As heavy rains upon recently plowed land are especially conducive to soil erosion, a study of the comparative frequency of heavy spring rains has been made. Figures 24 and 25 show the distribution of one type of such rain during April and May. Figure 24 shows that, in April, rainfalls of 2.5 inches, or more, in one day did not occur even once in the

twenty years studied, in most of northern Indiana, but occurred two or three times during the same period in much of south-central Indiana. Similarly, in May, one-day rains of 2.5 inches, or more, did not occur once in twenty years in two large northeastern and eastern areas, but

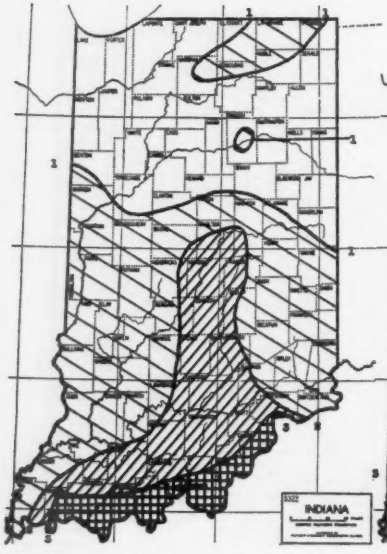


FIGURE 24.—*Distribution of April heavy rains*

Heavy rains designates rains of 2.5 inches in one day. Figures indicate times per twenty Aprils (1917-1936). Based on data from 86 stations.

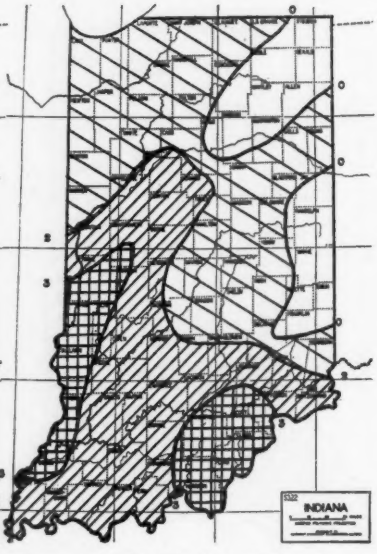


FIGURE 25.—*Distribution of May heavy rains*

Heavy rains designates rains of 2.5 inches in one day. Figures indicate times per 23 Mays (1914-1936). Based on data from 86 stations.

occurred two, three, or more times during the period in the southwestern third of the State. Of one-day rains of two inches, or more, maps, not published here, show that, in May, such rains are about five times as frequent in southwestern as in northeastern Indiana. In April, on the other hand, rainfalls of two or more inches in a day are somewhat more frequent in the extreme southeast than in the southwest, but, likewise, are almost unknown in a large northeastern section. The area of greatest frequency of such April rains "happens" to coincide roughly with the area of most extensive severe sheet erosion (Fig. 6).

Figure 26, showing the greatest rainfalls ever officially recorded in six consecutive days, presents further data of interest with respect to erosion, as the biggest rains of half a century sometimes accomplish as much erosion as all the lesser rains combined. This map is based on about fifty

years of record for many of the stations, but upon more than sixty years for a few.

Figure 26 reveals that the record six-day rain has not exceeded six inches in much of northeastern Indiana but has exceeded ten inches in a considerable share of southern Indiana. At the northeastern corner of the State the record is less than five inches, in contrast to that showing more than twelve inches in the extreme south-central part (13.2 at Jeffersonville). For the average of the northern third of the State, the record is less than six inches; for the central third, it is about 7.5 inches; and for the southern third, it is about ten inches. The seasonal distribution of these big rains is also of interest. Most of the biggest rains of northern Indiana have fallen during the warmest three months; more than three-fourths of those of southern Indiana have come during the cooler half of the year.

Whereas six inches of rain in six days is unknown in northeastern Indiana, such rains occur in one-ninth of the years at Indianapolis, and in one-sixth to one-fifth of the years along the Ohio River and the lower Wabash. Eight inches of rain in six days is almost unknown in the northern half of the State but such rainfalls occur in from three to eight per cent of the years in the southern area, roughly south of a line connecting Louisville, Kentucky, and Vincennes, Indiana (Fig. 1).

Hence, Figure 26 and the data it is based upon contribute further evidence that rainfall conditions in southern Indiana help to explain why the erosion is much greater there than in northern Indiana.

The striking regional contrasts in rainfall intensity in Indiana (Figs. 15 to 26) are clearly helpful in explaining the observed regional contrasts in erosion. As these contrasts are largely due to fundamental conditions, it is altogether probable that somewhat similar contrasts existed during, and previous to, the glacial period. Such regional contrasts in rainfall would help to explain the stoppage of the glacial advance, both because of the melting caused by the rain (in winter as well as in summer), and especially because of the greater ruggedness that the run-off from such rainfall produces.

The climatic contrasts which the foregoing maps reveal are practically all conducive to greater erosion in the south than in the north. Moreover, for most of the types of Indiana climatic data here mapped, the climatic conditions in the western part of any given latitude are more conducive to erosion than are those in the eastern part. Numerous other aspects of Indiana's climate have been studied. Of some of these, maps are published elsewhere,[8] and others will be published later. The same

[8] S. S. Visher: *Indiana regional contrasts in temperature and precipitation (31 climatic maps)*, Ind. Acad. Sci., Pr., vol. 45 (1936) p. 183-204.

general condition is revealed in practically all the maps that have been prepared—namely, considerable regional contrasts between northern, central, and southern Indiana, and appreciable differences between the eastern and the western parts of each of these zones. So far as erosion

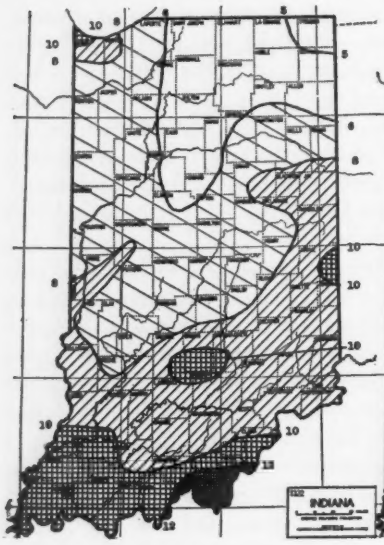


FIGURE 26.—*Distribution of greatest recorded six-day consecutive rainfall*

Figures indicate the amount of rainfall. Based on all available data from 86 stations. (The period averages about 45 years in length but for a few stations it exceeds 60 years.)

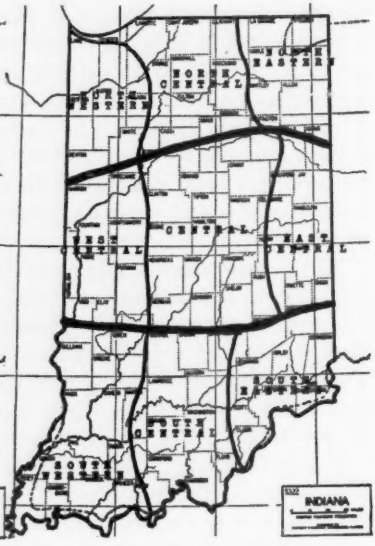


FIGURE 27.—*Indiana climatic regions and subregions*

is concerned, southern Indiana has more than northern Indiana in almost every aspect of the climate studied.

COMPARATIVE EROSION ACCORDING TO CLIMATIC REGIONS

The numerous climatic maps that have been prepared for Indiana warrant the division of the State into three major climatic regions—northern, central, and southern—each of which can appropriately be subdivided into an eastern, a central, and a western subdivision. Further minor subdivisions of a few of these subdivisions are possible. Figure 27 shows the climatic subdivisions now recognized and affords a summary of numerous climatic data. The boundaries between the various subdivisions are, of course, not sharp, but each of the subdivisions differs

from its neighbors in substantial respects, as may be seen by an examination of the thirteen climatic maps here presented.

Using the counties wholly or chiefly in the various regions shown by Figure 27 as the units, a comparison has been made between erosion and climate. The findings are presented in Table 4, which is comparable with Tables 1 to 3, except that this table is more detailed, as the subdivisions as well as the major types are included.

TABLE 4.—*Soil erosion and climatic regions*

Erosion type	Percentage								
	North east	North Central	North west	East Central	Central	West Central	South east	South Central	South west
Little erosion	46.6	65.8	45.4	71.9	78.3	62.2	38.1	19.8	43.5
Sheet erosion without gullying	34.0	15.1	6.0	1.1	4.5	7.0	0.3	0.8	3.0
Sheet erosion, moderate	5.3	26.5	16.4	23.1	20.9	38.2	26.5	67.6	45.7
Considerable sheet erosion	51.9	27.0	16.9	28.0	20.4	38.3	60.1	68.2	52.8
Severe sheet erosion	0.9	0.4	0.3	5.0	0.3	0.1	13.3	11.2	33.1
Considerable gullying	17.5	9.4	10.4	26.9	24.8	33.4	59.8	72.3	56.5
Severe gullying	2.0	1.0	6.0	0.5	0.9	12.3	11.8	49.6	32.4
Both gullying and sheet erosion considerable	17.5	11.1	10.1	22.5	17.0	32.3	60.2	77.6	56.2
Severe sheet erosion and severe gullying	1.0	0.5	0.3	0.5	0.0	0.0	7.4	9.4	11.4

A comparison of Table 4 with the earlier tables reveals that the correspondence between climate and erosion is at least as close as that between glaciation or relief and erosion. Indeed, except for the fact that northern Indiana has less land subject to little erosion than has central Indiana, there is a rather steady southward increase in erosion. The fact that the area of little erosion of the types here considered in northern Indiana is less than that in central Indiana is partly due to the rather widespread wind erosion in the northwest, and partly to the fact that some parts of the northeast have numerous morainic hills.

Southern Indiana is seen by Table 4 to be consistently more eroded than is central Indiana. It will be recalled that among the counties of considerable erosion (Figs. 5, 7, 9, 10) there was a less complete agreement for either the glacial or the relief data (Tables 1-3).

The erosional contrasts between the eastern, the central, and the western subdivisions of each of the climatic zones are not uniform, which fact proves that the climatic differences are not the only significant influences that affect the amount of erosion, for, of course, topography, soil, vegetal cover, and land-use are significant influences. The purpose of this paper is not to suggest that these influences are of minor importance, but, in-

stead, to offer evidence that a hitherto largely neglected influence is also of importance. The exceptions in Table 4 to the generalization suggested —that, in general, the climatic conditions of western Indiana are more conducive to run-off erosion (other things being equal) than are those of eastern Indiana, in similar latitudes—are partly due to the fact that wind erosion is more extensive in western than in eastern Indiana. As wind erosion is not included in Table 4, and as areas classified as eroded by the wind in Indiana generally are not classified as subject to water erosion, various western counties show in this table less erosion than is entirely correct.

LITHOLOGY, STRUCTURE, AND EROSION

From a geological point of view, discussion of erosion contrasts should also consider lithology and structure. For Indiana, no long discussion is required, as in most of the State the formations are almost horizontal and bed-rock is covered by glacial drift. Even in the approximately one-tenth of the State that was not glaciated, river alluvium and lacustrine deposits (in glacial marginal lakes) affect a considerable fraction of the area. Before deforestation, those parts of southern Indiana that had residual soil, possessed, in general, a comparatively deep and fairly uniform soil type, where mature. Indeed, detailed soil studies indicate that the *A* horizon (the topsoil) normally was deeper than in central or in northern Indiana, where the period since the Wisconsin glaciation has not been prolonged enough to permit the development of deep soil. Locally, in the unglaciated area and in parts of the Illinois drift sheet, bed-rock is exposed, and broad structural features are evident to the geologist. The escarpments, for example, of the Knobstone, or the outcrops of the Chester sandstone, are, however, feebly indicated by the soil erosion map analyzed. The reason for this is that the steeper rocky slopes have been less deforested and, hence, less subject to erosion, than are the less rugged areas. Indeed, one of the "inaccuracies" of the soil erosion survey was caused by the official directions that areas in forest or brush should be classed, in general, as only moderately eroded, even though portions of such slopes that had been cleared were promptly severely eroded.

In brief, efforts to relate regional erosion contrasts to lithology or structure in Indiana have not as yet been very fruitful. Locally, as in the Mitchell limestone belt with its many steep-sided sink holes, lithology is significant. But this discussion is of regional, not of local, contrasts.

The differences in soil, vegetation, and land use in Indiana are also largely local rather than regional, and do not require discussion here, although, in an exhaustive study, they merit attention, as there are

regional aspects for each. For example, the loessial soils of the southwest are especially subject to deep gullying; the sandy soils of the northwest are eroded most from the wind; and the heavy clays are especially subject to sheet wash wherever their upper layers have been loosened by freezing and thawing.

EROSION AND DIRECTION OF SLOPE

Although the rest of this paper is concerned with average regional contrasts, another type of extensive areal contrast in erosion is so conspicuous in southern Indiana that it merits a special section. This is the contrast in the erosion upon north-facing and that upon south-facing slopes.

Most south-facing slopes in southern Indiana are distinctly less steep than are the north-facing ones. Numerous measurements indicate that differences of five to ten degrees are common where the slopes are fairly steep. Even where both slopes are relatively gentle, south-facing slopes commonly have only about half the angular declivity of comparable north-facing slopes. In northern Indiana, however, slope contrasts of this sort usually are much less.

As these differences are extensive and distinctly significant to erosion, and are chiefly directly due to climatic differences, it is desirable that they be briefly discussed.

Four major climatic influences have co-operated to produce more active erosion upon south-facing than upon north-facing slopes in southern Indiana, which, in turn, has resulted in such slopes being cut down to a lower angle. (1) As remarked when Figure 16 (snow-cover) was discussed, the north-facing slopes remain snow-covered much longer than do the south-facing slopes, with the result that they are protected from part of the alternation of freezing and thawing and from run-off erosion. (2) The much more frequent thawings on south-facing slopes are especially significant in a region with as much winter rainfall as southern Indiana has, because the heaving of the soil, caused by the formation of ice crystals, often renders the top layer fluffy after the thaw, and easily carried away by run-off. The breaking of some of the shallower roots by the frost heaving is also of significance. Indeed, sometimes the alternation of freezing and thawing locally almost kills what had previously been a good growth of grass or winter wheat. (3) The rains of Indiana come largely with southerly or southeasterly winds and beat down slightly stronger on south-facing than on north-facing slopes. (4) Of special significance is the higher warm-season temperatures on south-facing slopes, which means greater evaporation and a greater water requirement for plants. As summer drouths are not infrequent,

the vegetative cover normally is much less dense on south-facing slopes than on north-facing slopes. Soil poorer on south-facing slopes than on north-facing slopes contributes to this poorer vegetation. As the amount of soil erosion depends, to no small degree, upon the character of the vegetative cover, this difference has been of real significance. A non-climatic factor also important is the fact that (5) despite lesser gradient and often drier soil, a larger share of the rainfall normally runs off from south-facing slopes than from north-facing slopes, because of the lesser vegetation and the less absorptive soil. Indeed, many of the steeper south-facing slopes of southern Indiana have relatively little topsoil, which is much more sponge-like than is most subsoil. Many of the steeper south-facing slopes in the unglaciated section of south-central Indiana have rock outcrops, whereas north-facing slopes of similar declivity and lithology have a thick mantle of soil.

Regional variations in these slope contrasts are due to differences in climate and glaciation. Such slope contrasts are most conspicuous in the unglaciated section of south-central Indiana but are notable in many parts of southern Indiana that were covered by the Illinoian glacier. In central Indiana, which was covered by the early Wisconsin ice-sheet, south and north slopes display much less average contrast. In northern Indiana, which was covered by the late Wisconsin ice sheet, slope contrasts due to the climatic influences just mentioned seldom are conspicuous.[9]

The length of time since glaciation has been significant in the development of these erosional contrasts, because, when forested (as most of Indiana was until a century ago), erosional changes were slow. North and south slopes differ more in southern than in central and northern Indiana, for climatic reasons also. Of special significance is the lesser snowfall in the south, with consequent greater differential in exposure on north and south slopes. Another factor is the shorter duration of constantly frozen soil in the south, and the less frequent alternation of freezing and thawing in northern Indiana. The much greater cool-season rainfall in the south is especially important in connection with soil loosening by thawing. The greater severity of drouths in southern than in northern Indiana is also important for its effect on the vegetative cover.

SUMMARY

The many data presented on the twenty-seven maps and four tables make a concise summary desirable. It is possible to present in a short

[9] Another kind of climate-induced slope-difference is present, however, on the numerous sand dunes that border Lake Michigan and those some distance south of the present lake, which bordered its post-glacial predecessor. These dunes are steepest on their south-facing slopes, in response to the stronger northerly winds, which were of chief importance in their development.

table only a few of the many facts, but some of the more significant ones for the three major regions are tabulated in highly generalized form in Table 5.

TABLE 5.—*Generalized summary of data presented on the maps*

	Northern Indiana	Central Indiana	Southern Indiana
Percentage of area subject to appreciable erosion	50	25	75
Percentage of area subject to considerable erosion	15	35	65
Percentage of area subject to severe erosion	0	0	10
Maximum local relief by counties (average, in feet)	100	150	300
Continuous freezing days	38	25	13
Snow-cover days	53	38	20
Cool-season rainfall (eight months, in inches)	21	25	30
Winter precipitation (inches)	6	8	11
Winter rains (2 inches in 2 days, per year)	0.5	4	12
Six-inch rains of four days (per cent of years)	0.3	5	9
Monthly totals of 10 inches (per cent of cooler half-years)	0	5	10
Excessive short rains (times per 30 winters)	0	2	7
April 2.5 inch rains (times per 20 years)	0.5	1.2	2.5
May 2.5 inch rains (times per 23 years)	1	1.8	2.8
Record six-day rains (inches)	5.5	7.5	10.0

CONCLUSIONS

There are marked regional contrasts in soil erosion in Indiana. Upon moderate slopes, on comparable soils, erosion is several times more rapid in the south than in the north. Upon steep slopes the differences are even greater. These contrasts are associated with contrasts in rainfall, for run-off is the great cause of soil erosion in Indiana. The one-third greater annual rainfall which the south receives is one cause of the contrasts in erosion, but the fact that, during the cooler eight months, the south receives nearly one-half more rainfall than does the north, and nearly twice as much in winter, is even more significant, as a larger percentage of cool-weather rain runs away. Differences in the length of time since glaciers covered the various parts of the State are also of significance, as the development of drainage run-off courses and related slopes required much time when the State was forested, as it nearly all was until about a century ago. Thus, the most recently glaciated area has, in general, on comparable slopes, less well developed drainage channels than have the areas glaciated earlier or not at all. The amount of local relief, a

special phase of the slope, is of further assistance in causing the observed regional contrasts in soil erosion. A climatic factor not previously investigated—namely, regional contrasts in rainfall intensity—clearly is also of great importance in causing the marked regional contrasts. Southern Indiana has from several to many times as many downpours of rainfall and consecutive days of heavy rain as does northern Indiana, and larger amounts of rainfall in its rainy spells. Intense rains yield much more run-off than do moderate rains, and, as the transporting power of run-off increases approximately as does the cube of the velocity, erosion increases enormously in places where heavy rains occur frequently, provided the soil is not protected by adequate vegetative cover.

Erosion greater in southern than in northern Indiana is, therefore, the result of a combination of influences, of which one of the more fundamental appears to be the differences in climate, which, in turn, are largely the results of relative location, for the differences in altitude are slight. Southern Indiana, because of its lower latitude, receives more intense insolation, with which are associated higher temperatures, greater atmospheric convection, and, hence, greater intensity of rainfall. It is also closer to the Gulf of Mexico, the significant source of Indiana's rainfall, and thus receives more rainfall. The location of northern Indiana nearer the central part of the continent helps to explain its lesser rainfall during the colder months, as continental interiors in mid-latitudes characteristically are relatively cold and dry in the winter.[10]

The causes of these climatic contrasts are of so fundamental a nature that it is altogether probable that they have long operated, and, hence, that the northern and southern parts of what is now Indiana long have differed climatically. As these climatic contrasts clearly produce erosion upon comparable slopes greater in the south than in the north, it follows that not only do these climatic differences help to explain the present differences in soil erosion, but they helped produce some of the larger physical features. For example, the southward extension of the Pleistocene ice sheets clearly was influenced not only by differences in temperature but also by differences in rainfall, and especially by the more rugged topography induced thereby.

Thus, in brief, a study of soil erosion in Indiana, a representative midwestern State, leads to the deduction that aspects of the climate hitherto little studied have contributed not only to regional contrasts in soil erosion, but to regional contrasts in the development of features of larger physiographic significance.

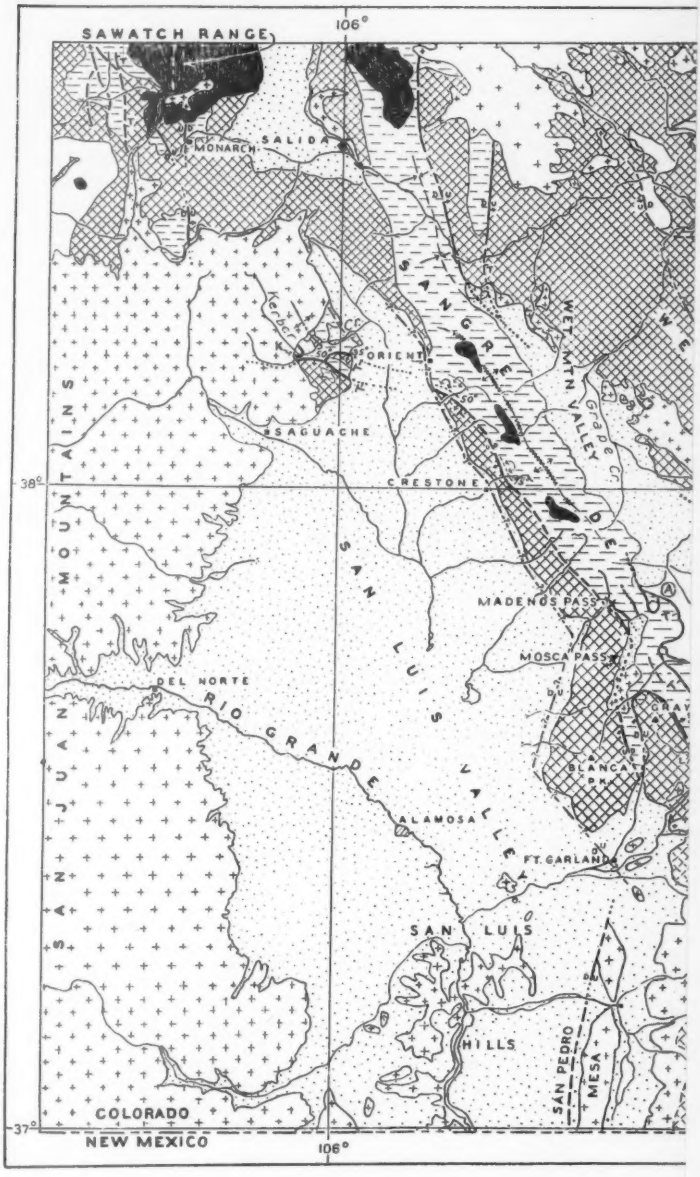
---

[10] S. S. Visher: *Climatic laws, ninety generalizations with numerous corollaries as to the geographic distribution of temperature, wind, moisture,* New York (1924).

Ellsworth Huntington and S. S. Visher: *Climatic changes, their nature and causes,* New Haven (1922).

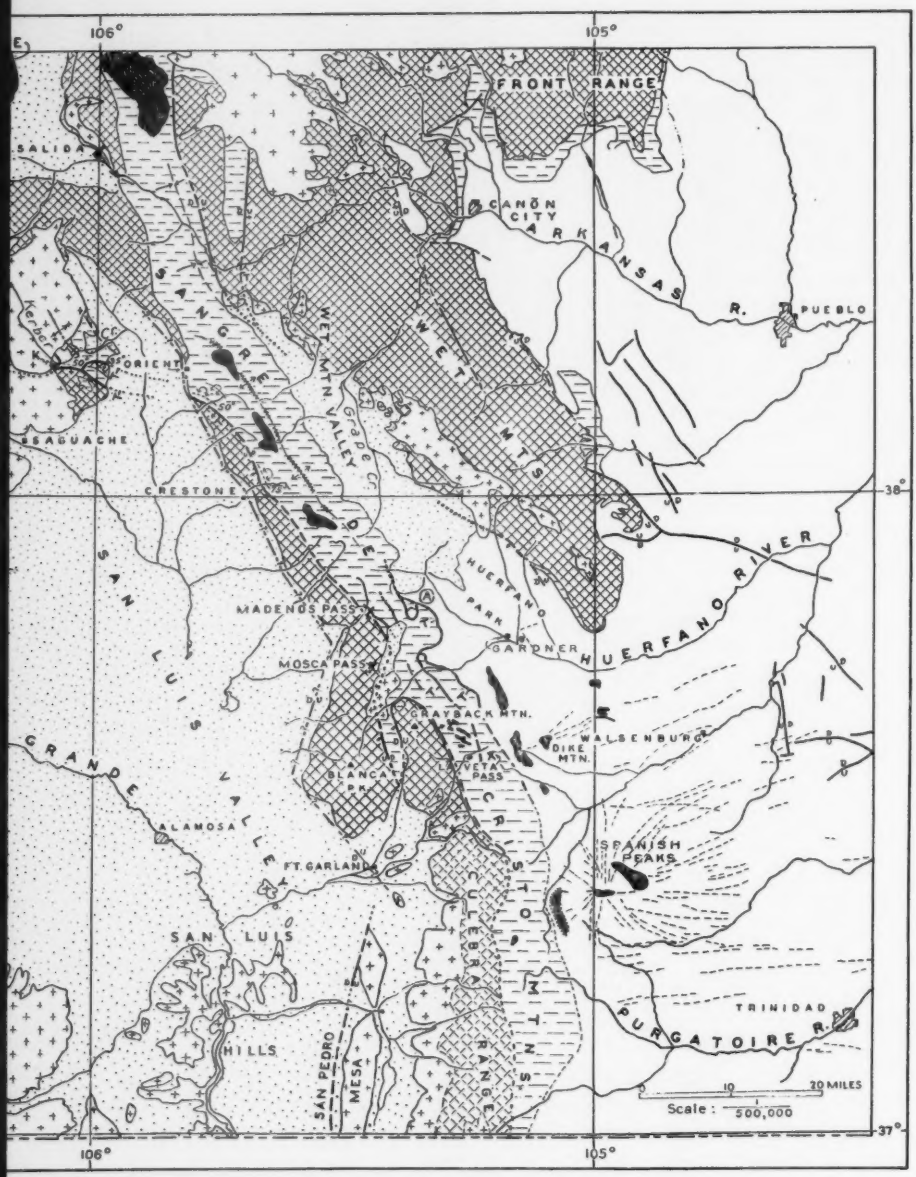
It now appears that, in order to understand fully the physical features of the earth, geologists should consider more than the character of the rocks, the diastrophic forces that have moved them, vulcanism, the influences of ancient incursions of glaciation, and the differences between the grosser climatic contrasts, such as those of arid as compared with humid regions, tropical as compared with sub-polar. The present study indicates that various lesser climatic contrasts are worthy of serious consideration, for example, moderate differences in intensity and seasonal distribution of rainfall, and in the frequency of freezing and thawings. Although such differences may be far less significant than are the forces commonly considered, nevertheless, they are important factors in many areas. Indeed, such effects of climate, acting slowly over long periods of time, apparently have been of considerable importance in affecting some of the more prominent geologic agencies. At any rate, in a representative area, Indiana, which is commonly considered to have unimportant regional climatic contrasts, climatic differences of considerable physiographic significance are revealed by detailed analyses.

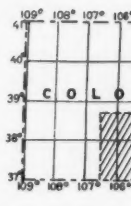
INDIANA UNIVERSITY, BLOOMINGTON, IND.
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JANUARY 4, 1937.
READ BEFORE THE GEOLOGICAL SOCIETY, DECEMBER 29, 1936.



GEOLOGIC MAP OF THE SANGRE DE CRISTO MOUNTAINS A

Based on geologic map of Colorado [with additional structural data] by th
with the Colorado Geological Survey Board and Colora





OF THE SANGRE DE CRISTO MOUNTAINS AND ADJACENT REGIONS, COLORADO

Colorado [with additional structural data] by the United States Geological Survey, in cooperation the Colorado Geological Survey Board and Colorado Metal Mining Fund, 1935.

EXPLANATION

Late Tertiary and Quaternary formations.

Tertiary volcanic rocks.

Intrusive bodies and dikes chiefly early Tertiary.

Mesozoic and early Tertiary formations.

Paleozic formations.

Pre-Cambrian massifs. Granite, schist and gneiss.

Normal and reverse faults. Arrows give direction of dip.

Overthrust fault with amount of dip.

K
Klippen at Kerber Creek.

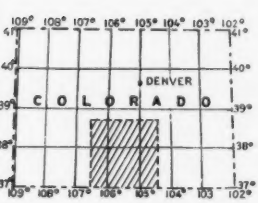
Anticline

Anticline with metamorphosed core (*partly injective or diapiric?*)

Overturned anticline.

Overturned syncline.

Ⓐ
Location of detailed map of Figure 4.



BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 48, PP. 931-976, 7 PLS., 4 FIGS. JULY 1, 1937

THRUSTING IN HUERFANO PARK, COLORADO, AND RELATED PROBLEMS OF OROGENY IN THE SANGRE DE CRISTO MOUNTAINS

BY W. S. BURBANK AND E. N. GODDARD

CONTENTS

	Page
Introduction	932
Previous geologic mapping	934
Geomorphic units of the region	935
Stratigraphy	938
Pre-Cambrian rocks	938
Ordovician beds	939
Devonian and Mississippian beds	939
Pennsylvanian and Permian formations	939
General features	939
Pennsylvanian and Permian formations undivided	941
Permian conglomeratic red beds	944
Other Permian rocks	945
Jurassic and Cretaceous sediments	945
Eocene sediments	946
General features	946
Physiographic and climatic conditions of Bridger time	947
Post-Eocene formations	949
Deformation in the marginal belt of the Sangre de Cristo Thrust	949
Definition and summary	949
North-central portion of the range	951
Huerfano Park region	955
General structure of deformed belt	955
Stages of deformation	957
Depth of marginal deformation	962
Deformation in the pre-Cambrian massifs	963
General features	963
Culebra massif	965
Sierra Blanca massif	969
Correlation of stages within the massifs	971
Igneous activity in relation to structural history	974
Conclusion	975

ILLUSTRATIONS

Figure	Page
1. Generalized and diagrammatic sections across the north-central part of the Sangre de Cristo Mountains	950
2. Geologic section from the Bruff Creek area west to the pre-Cambrian massif south of Madenos Pass	960

Figure Page
3. Diagrammatic illustration of plastic deformation in shale bodies near front of overthrust plates 964
4. Geologic section across Sierra Blanca and Culebra massifs to the edge of Huerfano Park 968

Plate Facing Page
1. Geologic map of the Sangre de Cristo Mountains and adjacent regions, Colorado 931
2. Stratigraphic columns of Paleozoic formations in the Sangre de Cristo region and at Ouray, Colorado 938
3. Stratigraphic column of Mesozoic and Tertiary formations of Huerfano Park and vicinity 946
4. Geologic sketch map of Huerfano Park, Colorado 948
5. Permian syncline 952
6. Views of structural features 953
7. Plane-table map and structure section of a lobate salient of the main thrust belt 956

INTRODUCTION

Huerfano Park, in south-central Colorado, is bounded on the west by the Sangre de Cristo Mountains and on the east by the southern end of the Wet Mountains (Pl. 1). At its south end, the park opens onto the plains near the termination of the bordering Wet Mountains uplift, but, at the north, it merges across a low divide into the Wet Mountain Valley, with which it is, in part, structurally continuous. Huerfano Park itself is of particular geologic interest, because there are Eocene sedimentary deposits of Green River and lower Bridger age, equivalents of which are not found elsewhere along the eastern front of the southern Rocky Mountains. Thrusts of the Sangre de Cristo Mountains intersect these remnants of Eocene sedimentary rocks along the western border of the park, and thus record some later stages of mountain-building that might, otherwise, be scarcely distinguishable from much earlier stages of the Laramide orogeny.

The region as a whole, including the San Luis Valley lying immediately west of the Sangre de Cristo Mountains, also affords excellent illustrations of different types of mountain and valley structure. The Sangre de Cristo Mountains include what is probably one of the most complexly folded sedimentary belts in Colorado. The Wet Mountain uplift, on the other hand, is a southward *en échelon* extension of the ancient Front Range highland of Colorado, a kind of structure commonly described in geologic literature as being repeatedly rejuvenated by forces that are expressed chiefly in differential vertical deformation. Huerfano Park and Wet Mountain Valley are downwarped intermontane basins or synclines, complicated by superimposed high-angle faults and bordering

thrusts. The San Luis Valley is a morphologic feature of later structural origin than the other units, and represents, in conjunction with the bordering pre-Cambrian massifs of the Sangre de Cristo Mountains, a type of basin and range structure. This has been produced by subsidence and disintegration of the original crustal fold of the Sangre de Cristo Mountains and of the bordering broad domal uplift of the San Juan Mountains.

The structural origin of the southern Rocky Mountains, of which this region is a cross-section, has long been a matter of debate. To a typical section across the entire width of the mountains in northern Colorado, R. T. Chamberlin [1] applied the theory of lateral compression, and computed the theoretical depth of folding. In a cross-section near the southern end of the Sangre de Cristo Mountains at Mt. Trinchera, Shepard [2] described shortening which he considered of greater relative magnitude than that shown in northern Colorado. Lee [3] maintained, on the other hand, that lateral compression was not in harmony with many of the observed facts, and that field observations indicated that the mountains were formed chiefly by vertical uplift. Recent studies in different parts of Colorado have revealed, however, the existence of a number of thrust-faults of considerable magnitude,[4] which is in disagreement with one of Lee's principal contentions. The present paper, which describes large thrust faults as well as vertical uplifts in southern Colorado, contributes further data for the solution of the problem.

The incentive for a study of the Sangre de Cristo region began with the discovery of a major overthrust zone in the Kerber Creek district in 1927, and was further increased by the desire for more accurate representation of the Sangre de Cristo Mountains on a proposed new geologic map of Colorado. For the purpose of tracing the Kerber Creek fault zone and to determine the general structure, several trips were made into the range by the senior author, mostly in the company of Dr. B. S. Butler, but also, in part, during field work in the Bonanza mining district on Kerber Creek.

The final field work in Huerfano Park was undertaken by the writers

[1] R. T. Chamberlin: *The building of the Colorado Rockies*, Jour. Geol., vol. 27 (1919) p. 145-164, 225-251.

[2] F. P. Shepard: *Indications of important horizontal compression in the Colorado Rockies*, Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 403-408.

R. T. Chamberlin: *On the crustal shortening of the Colorado Rockies*, Am. Jour. Sci., 5th ser., vol. 6 (1923) p. 215-221.

[3] W. T. Lee: *Building of the southern Rocky Mountains*, Geol. Soc. Am., Bull., vol. 34 (1923) p. 285-300.

[4] T. S. Lovering: *Williams thrust-fault* (abstract), Geol. Soc. Am., Bull., vol. 39 (1928) p. 173.

W. S. Burbank: *Preliminary report on some thrust-faulting involving the Paleozoic formations in the Kerber Creek region of southern Colorado* (abstract), Geol. Soc. Am., Bull., vol. 39 (1928) p. 172-173.

Also a number of other faults in central Colorado, mapped in part by B. S. Butler, Charles H. Behre, Jr., J. W. Vanderwilt, and others: *Geologic Map of Colorado*, U. S. Geol. Surv. (1935).

in the fall of 1931. Inadequacy of topographic maps and the complexity of structure permitted detailed mapping of only a small area along the west side of the park. In the park the remnant of the original belt of deformation marginal to the Sangre de Cristo fold is wider than elsewhere, and the relations of the different stages of deformation to the Tertiary formations are well shown. The studies in this park thus formed a basis for clearer understanding of the less detailed studies in the mountains as a whole.

The plan of this paper is to present, first, a brief review of the geomorphic and stratigraphic units of the region, followed by an analysis of the structures exposed in the marginal belt of the Sangre de Cristo thrust zone and, finally, of those in the pre-Cambrian massifs of the range. Suggested correlations of the different stages of deformation that were recognized in these two units are also briefly discussed, and their possible significance in the general problem of orogeny in the Colorado Rockies incidentally developed.

The work suggests that, within the zone of geologic observation, both tangential thrusting and vertical uplift have been involved in the building of the mountains, and that differential vertical movements became predominant during the later stages of deformation. The effects of these later stages, because of their more recent control of erosion and topographic form, have to some extent masked, and even completely obscured, the effects of earlier stages of tangential thrusting.

The writers conclude that the structure in the Sangre de Cristo Mountains does not as a unit permit precise analysis, either solely on the basis of the lateral compression theory of Chamberlin, or by the application of isostatic principles to vertical uplift, as treated by Lee. The lateral-compression theory, however, is applied in this paper to determine the depth of folding in a selected portion of the zone of thrusting and faulting that involves only the shallower phases of deformation. No attempt is made to apply or to derive any general theory as to the ultimate causes of the mountain-building.

The writers wish to acknowledge valuable suggestions in the course of the work by their colleagues in Colorado and especially to Dr. B. S. Butler, with whom the first reconnaissance studies of the mountains were undertaken. They are also indebted to G. F. Loughlin, W. W. Rubey, and C. P. Ross, of the United States Geological Survey, for reading and helpful criticism of the original manuscript.

PREVIOUS GEOLOGIC MAPPING

Little detailed geologic mapping has been done in the Sangre de Cristo Mountains. The *Geological and Geographical Atlas of Colorado* and the reports published as a result of the territorial surveys of 1873 to 1876,

under F. V. Hayden,[5] contain the only original maps of the geology of the range as a whole, and the reconnaissance nature of this early work permitted only very generalized representation. The text and sections fail to convey a clear idea of the structure, inasmuch as the central core of the range is shown as metamorphic granite. Folios of the *Geologic Atlas of the United States* have been published in the quadrangles immediately southeast and east of Huerfano Park. The Walsenburg quadrangle[6] includes a portion of the southern end of the Wet Mountains and adjacent plains, and the Spanish Peaks quadrangle[7] includes the intrusive bodies lying east of the Sangre de Cristo Mountains, but none of the range itself. A map of the Grayback mining district, south of Huerfano Park, has been published by the Colorado State Geological Survey.[8] Other stratigraphic and structural studies will receive attention on subsequent pages.

Doubtless, the lack of good topographic maps accounts in a large measure for the almost complete lack of comprehensive areal surveys of the mountains. The known mineral resources of the area, however, have furnished little incentive for either governmental or commercial surveys.

GEOMORPHIC UNITS OF THE REGION

That the reader may more readily appreciate the relation of the structures to be described to the general geologic features of the southern Rocky Mountains, physiographic and geologic units of the Sangre de Cristo region (Pl. 1) will be briefly outlined.

The Sangre de Cristo Mountains extend in a south-southeasterly direction from the Arkansas River in Colorado into New Mexico. The northern division of these mountains in Colorado, extending to the Blanca Peak massif or Sierra Blanca, has been shown on some maps as the Sangre de Cristo Range, whereas the southern division was called the Culebra Range. In this paper, the term Sangre de Cristo Mountains is used to include both divisions. Most of the work was within the northern division. The range is characterized by its small width, from less than 10 to about 20 miles, and by its regular arc-like form, convex toward the east. The average altitude of the central peaks is a little more than 12,000 feet, and their regularity is broken by only a few comparatively

[5] F. V. Hayden: *Geological and geographical atlas of Colorado and portions of adjacent territory*, U. S. Geol. Geog. Surv. Terr (1877) 827 pages.

F. M. Endlich: *Geology of Section B, San Luis division*, 7th Ann. Rept., U. S. Geol. Geog. Surv. Terr., (1874) p. 323-334; *Sangre de Cristo Range and the Huerfano region*, U. S. Geol. Geog. Surv. Terr., 9th Ann. Rept. (1877) p. 108-139.

[6] R. C. Hills: *Description of the Walsenburg quadrangle*, U. S. Geol. Surv., Geol. Atlas, folio no. 68 (1900) 6 pages.

[7] R. C. Hills: *Description of the Spanish Peaks quadrangle*, U. S. Geol. Surv., Geol. Atlas, folio no. 71 (1901) 7 pages.

[8] H. B. Patton and others: *Geology of the Grayback mining district, Costilla County, Colorado*, Colo. Geol. Surv., Bull. 2 (1910) 111 pages.

low passes. The range is bounded on the west by the San Luis Valley depression, from which it is apparently separated by profound fault lines, as this linear physiographic boundary cuts across some of the principal structural divisions of the range. On the east, the range is bounded by the Wet Mountain Valley and the basin of the Huerfano River. The boundary between the east flank of the range and the Wet Mountain Valley is, likewise, surprisingly linear but is apparently controlled by uniform easterly dips of the sedimentary formations that form the flank. North of Wet Mountain Valley, the east side of the range is bounded by a prominent fault, which probably continues southeastward beneath the alluvial deposits of the valley and joins with a similar steep fault which separates Huerfano Park from the west side of the southern Wet, or Greenhorn, Mountains. The physiography of the east boundary, from the head of Huerfano Park south to the State line, is controlled by the thrust faults and upturned beds that border the belt of marginal deformation, and is especially characterized by prominent hogbacks of the steeply tilted resistant beds. The folds, minor flexures, and igneous intrusions incident to the development of the range extend some distance from the mountain front eastward into the Great Plains province, but the folds are thought to affect chiefly the comparatively shallow blanket of sedimentary beds overlying the basement rocks.

The southern end of the Sawatch Range, shown in the northwest corner of Plate 1, is nearly in line with the northernmost physiographic continuation of the Sangre de Cristo Mountains, and in some publications the two ranges are spoken of as parts of the same line of deformation separated by a deflection, but, as will be shown presently, the two ranges are structurally distinct. As followed northward, the belt of greatest thrusting in the Sangre de Cristo Mountains swings westward, diagonally across the western part of the mountains, in an arc much sharper than that formed by the northern end, and meets the southeasterly continuation of the Sawatch line of folds in an obtuse angle at Kerber Creek, near the head of the San Luis Valley. The Sangre de Cristo arc thus joins the folds and thrusts of the Sawatch Range by linkage, as defined by Suess. The direction of thrusting along the Sangre de Cristo arc at this position is north-northeastward, whereas that of the greatest thrusts of the Sawatch Range is westward.

The Wet Mountains form a southward continuation *en échelon* of the Front Range of Colorado and, like the latter range, are composed chiefly of pre-Cambrian rocks. In the southern part of the Wet Mountains, where the pre-Cambrian core is capped with lavas, the highest summits are about 11,500 feet above sea level. This portion of the range is called the Greenhorn Mountains. Where exposed, the western border of the Wet

Mountains is characterized by steep faults or by sharply upturned beds of the bordering sedimentary formations. The eastern border, however, is locally characterized by moderate overturning or overthrusting of the formations toward the east.[9] The uplift is, therefore, asymmetrical but is apparently reversed with respect to the asymmetry of the Front Range, where the greatest overthrusts have been found along the west side.[10]

The topographically rejuvenated portion of the ancient Wet Mountain arch ends abruptly at the edge of the Great Plains, but, as this arch is at least as old as Paleozoic in origin, it probably extends southeastward, beneath Mesozoic deposits, toward the Las Animas uplift of southeastern Colorado and the Wichita Mountains of Texas and Oklahoma.

From a point near the north end of the Sangre de Cristo Mountains the San Luis Valley extends southward into New Mexico, with a total length of about 150 miles and a maximum width of about 50 miles. As pointed out by Siebenthal,[11] the valley floor is so nearly level that one travelling over it scarcely realizes that it has been formed by the coalescing of alluvial fans from the surrounding mountains. The western border of the valley, as shown in Plate 1, has an origin and physiographic expression entirely different from that of its eastern border. The eastern hills of the San Juan Mountains have become gradually submerged beneath the alluvial accumulations of the valley until their indented boundary resembles that of a drowned coast. Both this feature and the presence of lava-capped hills, which rise above the alluvium in the southern part of the valley, show that the valley is structurally an eastern part of the San Juan Mountain block. The younger Tertiary volcanic deposits and interbedded gravels dip 3 to 5 degrees eastward beneath the valley, and it appears that the San Juan block was tilted eastward while they were being deposited. According to this interpretation, the total thickness of sediments and lavas beneath the structural axis of the valley, which is close to the Sangre de Cristo Mountains, must be very great.

The Wet Mountain Valley, like the San Luis Valley, is floored by alluvial deposits from the bordering ranges, but on its east side and at

[9] C. W. Washburne: *The Canon City coal field, Colorado*, U. S. Geol. Surv., Bull 381 (1908) p. 342, pl. 18.

[10] C. W. Washburne: *The South Park coal field, Colorado*, U. S. Geol. Surv., Bull. 381 (1908) p. 307-316, pl. 16.

T. S. Lovering: *Geology and ore deposits of the Breckenridge mining district, Colorado*, U. S. Geol. Surv., Prof. Pap. 176 (1934) p. 17-18.

T. S. Lovering: *Geology and ore deposits of the Montezuma quadrangle, Colorado*, U. S. Geol. Surv., Prof. Pap. 178 (1935) p. 47.

C. H. Behre, Jr., I. T. Schwade, and R. M. Dreyer: *Bedrock geology of northern South Park* (abstract), Geol. Soc. Am., Pr. 1934 (1935) p. 66-67.

J. T. Stark, and A. D. Hoagland: *Bedrock geology of central South Park* (abstract), Geol. Soc. Am., Pr. 1934 (1935) p. 109.

[11] C. E. Siebenthal: *Geology and water resources of the San Luis Valley, Colorado*, U. S. Geol. Surv., W. S. Pap. 240 (1910) p. 9-12.

both ends, late Tertiary volcanic and sedimentary deposits pass beneath the alluvium. The early Tertiary deposits of Huerfano Park pass northward beneath the late Tertiary deposits that are exposed at the head of the park, and available evidence as a whole implies that Huerfano Park and Wet Mountain Valley together have a synclinal structure that has been dislocated by faulting and partly concealed, especially in the western part, by alluvium. The longitudinally faulted synclinal structure is similar to that of North, Middle, and South Parks, of Colorado, which form a linear group, and it may be regarded as a fourth member of the group. The early Tertiary rocks at the southern end of the valley, near the head of Huerfano Park, were tilted and broken, probably in late Eocene time, by the fault that extends from the head of the park northward beneath the alluvial deposits of the Wet Mountain Valley. The later Tertiary sedimentary and volcanic rocks were then deposited, overlapping the partly eroded northern end of the Huerfano syncline and covering the extension of the fault line. These beds were later subjected to a minor degree of synclinal folding and perhaps renewed movements along the fault line, thus forming the basin that is now filled with alluvium.

STRATIGRAPHY

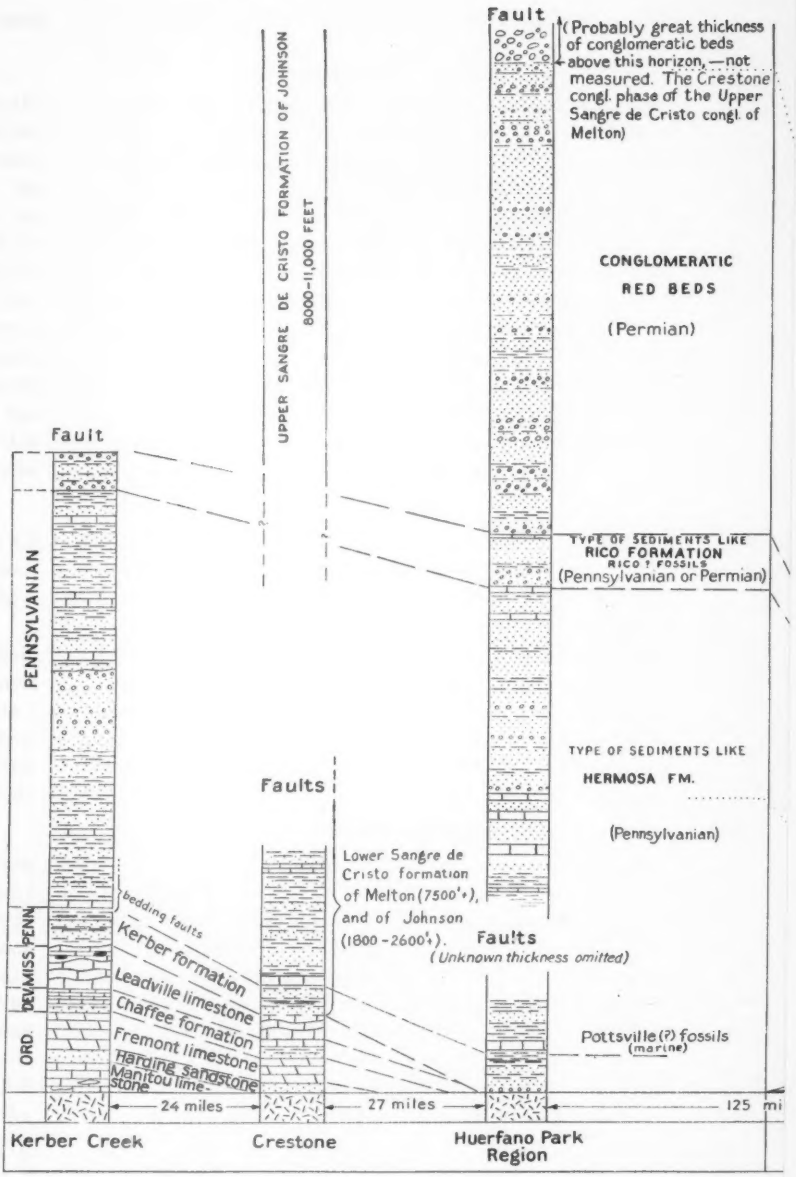
PRE-CAMBRIAN ROCKS

Only the briefest reference to the pre-Cambrian rocks is made in this paper, for neither their varieties nor their distribution and structure have received adequate study. Schist, gneiss, and pegmatite form the greater part of the massifs along the west border of the Sangre de Cristo Mountains, but there are also local bodies of granite and diorite. The western portion of the Blanca Peak group is said to be composed chiefly of coarse-grained granite,[12] but gneissic and schistose rocks are exposed on the east side of the peak, near the head of the Huerfano River and along the tributaries of Ute Creek, which drain into the San Luis Valley from the southeast side of the peak. At Mosca Pass, about 10 miles north of Blanca Peak, the principal rock is hornblende gneiss highly injected by granite and pegmatite.

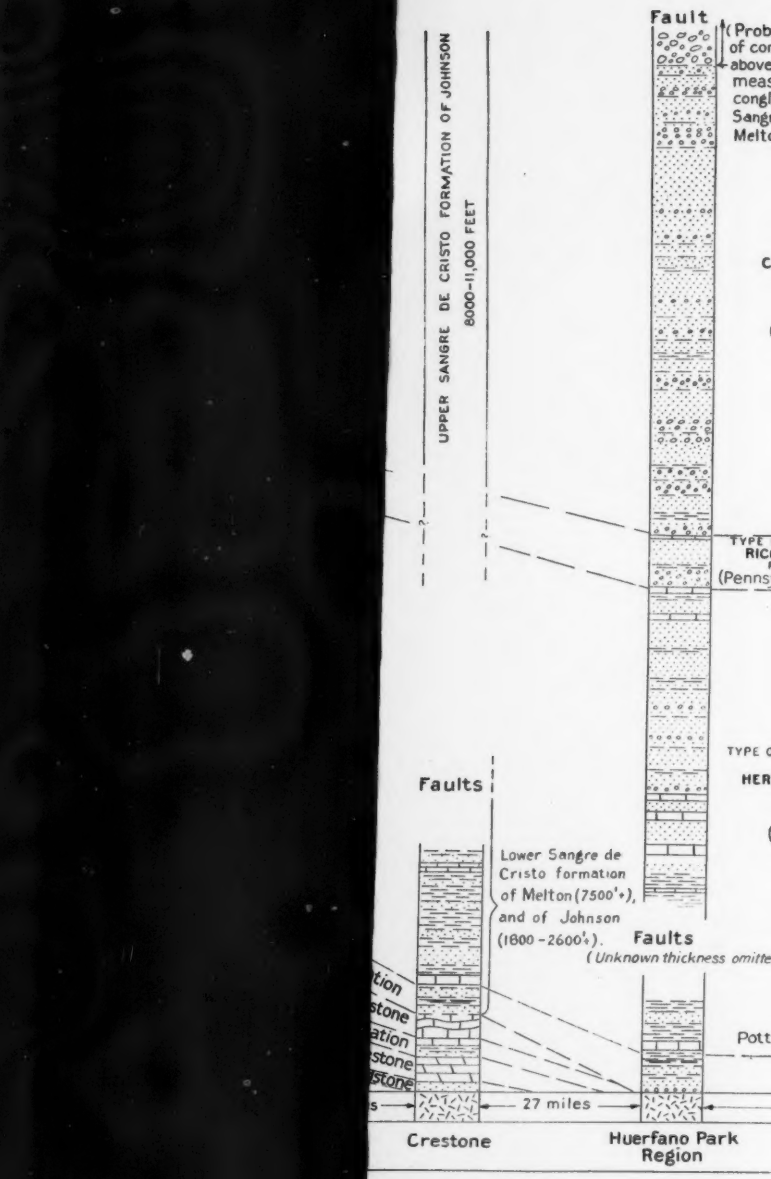
Farther north, near Crestone, more basic dioritic and hornblendic facies of pre-Cambrian igneous rocks are prominent.

In the vicinity of Mosca Pass the strike of the gneiss is N. 75° E. to E.-W., and the dip is steep to the south. The rocks there are cut by a set of slickensided shear planes, which trend N. 30° E., essentially parallel to the local contact with the Carboniferous rocks and dip at low angles both to the east and to the west. According to Siebenthal, a northeast-south-

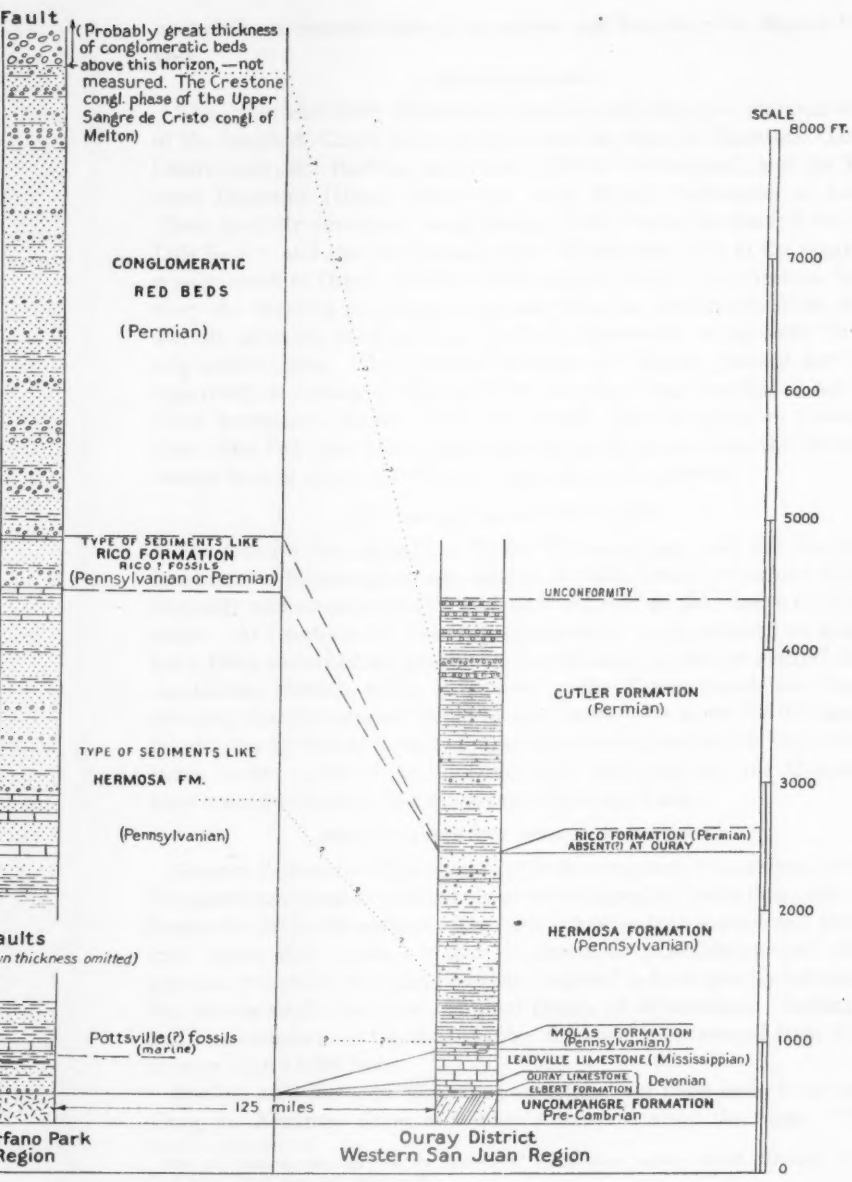
[12] C. E. Siebenthal: *op. cit.*, p. 36.



STRATIGRAPHIC COLUMNS OF PALEOZOIC FORMATIONS IN THE SA
COLORADO



RAPHIC COLUMNS OF PALEOZOIC FORMAT



ZOIC FORMATIONS IN THE SANGRE DE CRISTO REGION AND AT OURAY, COLORADO

west strike is characteristic of the schists and fissures in the Blanca Peak area.

ORDOVICIAN BEDS

The Ordovician beds, which immediately overlie the pre-Cambrian rocks of the Sangre de Cristo Mountains, include the Manitou limestone (Lower Ordovician), the Harding sandstone (Middle Ordovician), and the Fremont limestone (Upper Ordovician, with Middle Ordovician at base). These are fully developed along Kerber Creek,[13] near the head of the San Luis Valley, and also are exposed along the northern part of the range, to a point south of Orient (Kerber Creek section, Pl. 2). At Crestone, however, the Harding sandstone, with less than its normal thickness, rests directly upon the pre-Cambrian, indicating proximity to an early Paleozoic positive area. The Fremont limestone is, likewise, present here but apparently is reduced in thickness by shearing along the edge of the uplifted basement complex. It is not known how far south of Crestone these older Paleozoic beds extend, but west of Huerfano Park the Pennsylvanian beds in places rest directly upon the pre-Cambrian.

DEVONIAN AND MISSISSIPPIAN BEDS

The Chaffee formation,[14] of Upper Devonian age, and the Leadville limestone, of Mississippian age, overlie the Ordovician formations unconformably and co-extensively in the northern part of the Sangre de Cristo range. At Crestone the Leadville limestone is much sheared, so that it has a flaky and schistose structure. Its thickness is, thus, so reduced from the normal amount, which is exposed in the Kerber Creek and Orient districts, that its presence would be questioned were it not for the unmistakable recognition of those beds that underlie and overlie it in parts of the range farther north. Like the Ordovician, the Devonian and Mississippian beds are absent in the range near Huerfano Park.

PENNSYLVANIAN AND PERMIAN FORMATIONS

General Features.—The sequence of beds comprising Pennsylvanian and Permian formations forms the bulk of the sedimentary rocks that make the Sangre de Cristo Mountains. The beds comprise both marine and terrestrial facies, show much variation in degree of consolidation and other physical properties, and are, therefore, believed to have greatly influenced the development of shallow marginal phases of deformation. Estimates and measurements of the stratigraphic thickness have ranged from 2,000 to more than 13,000 feet.

Endlich used the term "Arkansas sandstone" for these beds, both those along the Arkansas River and those southward along the range. This

[13] W. S. Burbank: *Geology and ore deposits of the Bonanza mining district, Colorado,* U. S. Geol. Surv., Prof. Pap. 168 (1932) p. 19.

[14] Edwin Kirk: *The Devonian of Colorado,* Am. Jour. Sci., 5th ser., vol. 22 (1931) p. 220-240.

blanket term long ago fell into disuse, and, as "Arkansas" is now well established for a Devonian formation, it is not available for reviving for any divisions of this series. Hills[15] applied the term "Sangre de Cristo formation" to great thicknesses of conglomerates and sandstones in the part of the range that lies west of the Wet Mountains and the Spanish Peaks. He described the beds as being composed of "coarse reddish-brown conglomerate", or "conglomerate and sandstone", and stated that the formation had afforded remains of Upper Carboniferous fauna and flora. It may, therefore, be implied that he intended to include some, or all, of the beds containing upper Carboniferous marine fossils. This implication, however, does not agree with the emphasized lithologic character, and it may be that Hills was unaware that much of the series containing the fossils underlies the bulk of the coarse conglomerates. Melton[16] and Johnson[17] have applied the terms "Sangre de Cristo conglomerate" and "Sangre de Cristo formation" to a thickness of beds predominantly conglomeratic and sandy, but also including a lower division composed chiefly of sandstones, shales, and limestones, with some conglomerates. They both recognized the validity of a two-fold division, but apparently not upon the same basis. From both faunal evidence and lithologic succession, there seems to be justification for dividing the entire series into two major divisions, one predominantly of Pennsylvanian age, and the other of Permian age. In many, if not all, of the sections examined there is a distinct tectonic "disconformity" between the lower incompetent shaly beds of the Pennsylvanian and certain overlying more massive beds of either Pennsylvanian or Permian age, as partly shown by the sections illustrating faults in Plate 2. These apparent "disconformities" are believed to have resulted entirely from certain phases of deformation, described later, and are apparently unrelated to any major breaks in the sedimentary record. The writers, however, do not wish to take the position that important stratigraphic unconformities do not exist.

The general character and limits of the major divisions recognized are described later. The lower division, chiefly if not wholly, of Pennsylvanian age, consists of lithologic facies somewhat like Pennsylvanian formations elsewhere in Colorado. The upper division is composed mainly of conglomeratic red beds and is considered to be of Permian age. The writers wish to acknowledge the cooperation of Professor J. Harlan Johnson, of the Colorado School of Mines, and Dr. G. H. Girty, of the United

[15] R. C. Hills: *Description of the Walsenburg quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S. fol. no. 68 (1900) p. 1.

[16] F. A. Melton: *Correlation of Permo-Carboniferous red beds in southwestern Colorado and northern New Mexico*, Jour. Geol., vol. 33 (1925) p. 810-812.

[17] J. H. Johnson: *Contribution to the geology of the Sangre de Cristo Mountains of Colorado*, Colo. Sci. Soc., Pr., vol. 12 (1929) p. 6.

States Geological Survey, in preparing reports on fossil collections made during the field work.

Pennsylvanian and Permian Formations Undivided.—An inclusive name has not been applied to the lower of the major subdivisions, partly because it may be divisible into a number of formations when adequate data have been obtained. This opinion is based on reports on fossil collections and on comparison with the Pennsylvanian formations of the Mosquito Range, to the north, and of the San Juan region in southwestern Colorado.

The base of the section at Kerber Creek is composed of rather clean quartz sandstones with shale partings, overlain by alternating dark marine shales and limestones. The lower 200 feet of beds containing the sandstones, including plant-bearing shales, and impure coal seams, was named the Kerber formation.[18] The marine limestones and dark shales above were excluded because of apparent differences in conditions of deposition and because of the lack of a thick transition zone. Read [19] has more recently recognized the Pottsville age of a flora obtained perhaps from this same relative zone in the lower Weber (?) formation of the Mosquito Range, 65 to 70 miles north of Kerber Creek. A succession similar to that of Kerber Creek is evident east of Crestone and west of Huerfano Park, near Grayback Mountain, in the Sangre de Cristo Mountains. At the Grayback locality, however, the similarity in lithologic sequence is partly inferred, because the Pennsylvanian there rests upon the pre-Cambrian basement of the mountains, and not upon the Leadville limestone, as at the other localities; in addition, the carbonaceous shales are separated from the lowermost beds by faults. The basal beds at the north end of Grayback Mountain are composed in part of conglomeratic sandstones with quartz pebbles, and are less arkosic than are the higher sandstone beds of the Pennsylvanian in the vicinity. The carbonaceous shale and coaly beds are found locally along the west side of the mountain, in fault contact with the nearly vertical wall of granite, but are believed to represent part of the lower zone. The reason for this disturbed stratigraphic sequence is believed due to the uplifting of the mountain block, which locally sheared the incompetent beds from the flank of the mountain.

The lithologic zone lying immediately above the units just described is composed chiefly, in its lower part, of marine shales and limestones with thin interbedded sandstone layers. At Kerber Creek, a zone composed principally of shales and thin limestones was determined to have a thickness of about 300 feet, but, because of the persistence of bedding-plane faults along this zone, the significance of which was not fully recognized at

[18] W. S. Burbank: *Geology and ore deposits of the Bonanza mining district, Colorado*, U. S. Geol. Surv., Prof. Pap. 169 (1932) p. 13-14.

[19] C. B. Read: *A flora of Pottsville age from the Mosquito Range, Colorado*, U. S. Geol. Surv., Prof. Pap. 185-D (1934) p. 79-97.

the time of measurement, this figure may be somewhat in error for the total thickness. These shales grade upward into beds that are predominantly sandy and conglomeratic (Pl. 2). Along the Sangre de Cristo Mountains, exposures of this sequence were not sufficiently free from bedding faults or great thrusts to permit reliable measurements. Thus, the beds appear to have constituted a zone of incompetence during earlier stages of overthrusting, along which large thrust plates of the overlying more rigid beds were separated from, and flowed over, the basement.

Shales, sandstones, and thin limestones belonging to the lower zone appear also along the axis of the principal anticline of the range, between Orient and Crestone, but they are metamorphosed locally to schistose rocks. The beds in the frontal thrust plates east of Grayback Mountain indicate that possibly a much greater thickness of the lower zone becomes shaly and limy beneath the axis of the range and toward the south, as compared with the sequence farther northwest at Kerber Creek.

Two small collections, here listed, came from shales close to the base of the Pennsylvanian section, but above the basal sandstones and carbonaceous shales of the Kerber formation. According to Girty, they show a few rather tenuous relations with the fauna of the Morrow group of Arkansas, a well-known Pottsville fauna:

Lot 7136

East side Grayback Mountain near granite contact

Lophophyllum profundum ?
Derbya bennetti ?
Productus cora
Dielasma subspatulatum
Spirifer opimus
Spiriferina n. sp. ?
Composita gibbosa
Hustedia mormoni ?

Lower part of Pennsylvanian section, Kerber Creek

Crinoid fragments
Linguldiscina sp.
Schizophoria texana
Derbya crassa
Chonetes levis
Astartella concentrica
Myalina ?, n. sp.
Productus morrowensis var.
Productus cora
Composita subtilata
Acanthopecten carboniferous ?
Bucanopsis ? sp.
Pseudomonotis kansasensis ?
Phillipsia sangamonensis ?

Thus, these lower beds are believed to correspond with parts of the lower and middle zones of the Weber (?) formation of the Mosquito Range, as described by Johnson.[20]

Sequences of beds in the Sangre de Cristo Mountains, lying above the shaly sections described, were found to be interrupted in many places by faulted and crumpled zones that rendered the sections unsuitable for

[20] J. H. Johnson: *Paleozoic formations of the Mosquito Range, Colorado*, U. S. Geol. Surv., Prof. Pap. 185-B (1934) p. 30-31.

accurate measurement. One section, measured downward in stratigraphic sequence from the base of the Permian conglomeratic red beds west of Huerfano Park, consists of about 1900 feet of massive arkosic grits with thin intercalated shales and limestones (Pl. 2). Beneath these beds is a series containing a greater proportion of shale and arkosic limestone, but the proportion of bedding faults and contorted strata is, likewise, appreciably greater. The frontal thrust plates facing Huerfano Park (Pls. 4 and 6, fig. 1) are also believed to represent a part of this middle section of the Pennsylvanian, and are composed mainly of arkosic and conglomeratic sandstones with interbedded shales and limestones containing arkosic debris. A few fossil collections indicate that the fauna corresponds with that of the Hermosa formation of the western San Juan region. Johnson [21] has published a large faunal list from La Veta Pass, which he correlates with the Hermosa formation of southwestern Colorado and the Magdalena group of New Mexico.

Near the top of the Pennsylvanian section already mentioned, there are, interbedded in the conglomerates and grits, a few thin limestones and shales containing a marine fauna, and reddish or dark-colored shales containing plant remains. According to Girty, the few fossils in the marine shales and limestones may represent the fauna of the Permian Rico formation, which overlies the Hermosa in the western San Juan Mountains.

The general correspondence in lithologic succession and the suggested correlations are shown in Plate 2. That there should exist such a correspondence between the marine Carboniferous rocks in the Sangre de Cristo trough and those along the eastern border of the Carboniferous basin of southwestern Colorado may be deduced from the Paleozoic history and alignment of landmasses. The Paleozoic San Luis Range was probably an assymetrical uplift like the early Tertiary range, with a higher and steeper eastern front facing the Sangre de Cristo trough. The back slope of the uplift descended more gently beneath the Paleozoic basin of southwstern Colorado.

Relative changes in coarseness of debris from this landmass seem to be recorded just before, and during, the deposition of the Rico and Cutler formations, on the west, and the Permian conglomeratic red beds, on the east. The uppermost Hermosa faunas are in shales or limestones interbedded with arkosic grits or conglomerates. On both sides of the range the Rico faunas overlie, and are interbedded with, coarse clastic or conglomerate debris. In the section measured west of Huerfano Park the boulders within the Rico (?) equivalent reach a maximum diameter of about one foot, but few attain that size in the western San Juan.

[21] J. H. Johnson: *Contribution to the geology of the Sangre de Cristo Mountains of Colorado*, Colo. Sci. Soc., Pr., vol. 12 (1929) p. 7-9.

Thus, the beds confined strictly to the lower divisions of the Carboniferous in the Sangre de Cristo Mountains seem to show only moderate differences, such as that in the arkosic nature of some clastic limestones, from typical Pennsylvanian successions in central and western Colorado. Toward the south, these divisions presumably grade into the Magdalena group of New Mexico, which consists largely of limestone.

Permian Conglomeratic Red Beds.—The conglomeratic red beds of the Sangre de Cristo region have received much attention in geologic literature of the region, with special emphasis upon the origin of the beds and the extreme coarseness of some of the debris. In this second respect, and also with regard to their great thickness, there is no close counterpart in the Upper Carboniferous in other parts of Colorado. The upper conglomeratic parts of the Maroon formation of the Anthracite and Crested Butte quadrangles,[22] and of the Kangaroo formation of Crawford, in the nearby Monarch district [23] in the southern end of the Sawatch Range, correspond most closely in lithologic character. Melton [24] has given much attention and study to the distribution and nature of the upper conglomerates of the Sangre de Cristo Range, and, according to his data, they attain greatest thickness and coarseness in the vicinity of Crestone (13,000 feet thick) and decrease in thickness and coarseness both northward to the valley of the Arkansas (4300 feet) and southward to the Culebra, or southern, part of the Sangre de Cristo Range (6,000 feet). Melton concludes that they are derived from a Paleozoic mountain range that lay in approximately the present position of the San Luis Valley, and he, accordingly, names this the Paleozoic San Luis Range.[25] This range extended northwestward, as does the present Sangre de Cristo arc of thrusting, toward the Paleozoic Uncompahgre highland of western Colorado. The coarsest boulders in the beds are commonly several feet in diameter, and blocks 15 or 20 feet across have been reported. The Crestone conglomerate phase of Melton occupies the highest stratigraphic position in the central part of the range near Crestone.

There is general agreement among students of these beds that they were derived in the form of coalescing fans from the front of a high landmass facing a Paleozoic trough of the sea between the San Luis Range and the Wet Mountain landmass. In later epochs the sea had retreated from

[22] C. W. Cross and G. H. Eldridge: *Description of the Anthracite and Crested Butte quadrangles*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 9 (1894) p. 6.

[23] R. D. Crawford: *Geology and ore deposits of the Monarch and Tomichi districts, Colorado*, Colo. Geol. Surv., Bull. 4 (1913) p. 70-73.

[24] F. A. Melton: *Correlation of Permo-Carboniferous red beds in southwestern Colorado and northern New Mexico*, Jour. Geol., vol. 33, no. 8 (1925) p. 807-815.

[25] F. A. Melton: *The ancestral Rocky Mountains of Colorado and New Mexico*, Jour. Geol., vol. 33, no. 1 (1925) p. 84-89.

this trough, and the deposits were mostly terrestrial fans. The original width of deposits at the narrowest part of the trough, near the center of the range, may be approximately estimated from the amount of shortening by thrusting and folding during the Laramide orogeny. An original width of 15 or 18 miles, or perhaps even greater, is possible, for it is not certain whether or not a supposed overlap of the conglomerate upon the basement south of Crestone represents the real front of the western Paleozoic landmass.

Formation names have not been applied to the Permian and Pennsylvanian, in this paper, because of the inability of the writers to recognize the limits of formation names previously applied. Contradictions in definition of the units seem to have resulted because of failure to recognize the presence of thrusts along incompetent zones and the imbrication of beds in certain parts of the range. The Permian conglomeratic red beds as shown in Plate 2 are defined as comprising beds from the top of the uppermost marine shales and limestones with an invertebrate fauna of Rico (Permian) aspect to the top of the conglomerate beds exposed in the range. At the top is included the Crestone conglomerate phase of Melton. The entire division extends along the range, from the vicinity of the Monarch district, where the Kangaroo formation of Crawford seems to be its lithologic equivalent, to at least the state line in southern Colorado, a distance of 125 miles.

Other Permian Rocks.—At the eastern edge of the mountains, extending south from the vicinity of Huerfano Park, a series of red beds immediately underlies the Morrison formation. These are chiefly shales, fine sandstones, and red limy beds, with some conglomeratic lenses; they have an exposed thickness of at least 1,000 feet. Because of the intervention of bedding faults and thrusts, this series cannot be directly correlated with the Permian beds of the main range. It seems possible that they may interfinger toward the west with the more conglomeratic beds of the mountains, but could represent a younger epoch of sedimentation. In contrast to the rigidity of the conglomeratic formations, these beds were easily deformed and thus have an important role in the more plastic phases of deformation along the eastern edge of the mountains.

JURASSIC AND CRETACEOUS SEDIMENTS

Jurassic and Cretaceous sediments of Huerfano Park have not been fully described in earlier writings, but, from brief examination of the complete section, supplemented by a few measurements, it is apparent that the formations correspond approximately in lithology and thicknesses to those of the Walsenburg quadrangle.[26] The reader is, therefore, re-

[26] R. C. Hills: *Description of the Walsenburg quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 68 (1900) p. 1-3.

ferred to Plate 3 for the general lithology. Where necessary to conform with current usage, the earlier nomenclature of Hills has been modified.

Beginning with the rocks of Benton age, the beds extending to the top of the Pierre shale form an unbroken sequence of marine formations of Upper Cretaceous age. In the Walsenburg quadrangle, and farther south along the east border of the range, the Trinidad sandstone overlies the Pierre shale and constitutes the last phase of Cretaceous marine sedimentation, but this formation has not been identified in Huerfano Park. The Laramie formation of the Walsenburg quadrangle, as defined by Hills,[27] included beds that were later assigned by Lee [28] to the Vermejo and Raton formations. The Vermejo formation is recognized as definitely of Cretaceous age, but the Raton is classified by some geologists as questionably Cretaceous and by others as Eocene. The United States Geological Survey classifies it as Eocene.

At certain places in Huerfano Park the Pierre shale is overlain unconformably by the Poison Canyon formation, thus indicating a considerable hiatus that represents the latest part of Cretaceous and possibly the earliest part of Tertiary time. Although Hills [29] reports that his Laramie (Vermejo and Raton formations of current nomenclature) is prominently exposed just north of the Huerfano River, the writers failed either to find or to recognize any outcrops of these younger Cretaceous beds. The Trinidad sandstone and the Vermejo formation are believed to be absent at their proper stratigraphic position in a large part of the area north of the Huerfano River.

EOCENE SEDIMENTS

General Features.—Hills [30] divided the Eocene sediments of the Huerfano Basin into three formations: the Poison Canyon, the Cuchara, and the Huerfano. To these should possibly be added the Raton formation, which, in the present state of knowledge, is considered to be beneath the Poison Canyon, and which is assigned to the Eocene by the United States Geological Survey. The general lithology of these formations is shown in Plate 3, and their distribution in Plate 4.

Knowlton [31] has shown that the flora of the Raton formation has close relationship with that of the Denver formation, the Dawson arkose, and the Fort Union formation. The relation between the Poison Canyon and the Raton formations is, in some respects, similar to that between the

[27] *Op. cit.*, p. 2.

[28] W. T. Lee: *Geology and paleontology of the Raton Mesa and other regions in Colorado and New Mexico*, U. S. Geol. Surv., Prof. Pap. 101 (1917) p. 51-61.

[29] R. C. Hills: *The recently discovered Tertiary beds of the Huerfano River basin*, Colo. Sci. Soc., Pr., vol. 3 (1889) p. 156.

[30] R. C. Hills: *Remarks on the classification of the Huerfano Eocene*, Colo. Sci. Soc., Pr., vol. 4 (1891) p. 7-9.

[31] F. H. Knowlton: *The flora of the Denver and associated formations of Colorado*, U. S. Geol. Surv., Prof. Pap. 155 (1930) p. 9, 10.

PERIOD	FORMATION		THICKNESS IN FEET	GENERAL CHARACTER
PLIOCENE OR PLEISTOCENE	Terrace gravels			
EOCENE	Huerfano fm.		2300-3500	Marls, clays, soft sandstones and shales, predominantly red, but in part gray, yellow green and purple.
EOCENE	Cuchara fm.		300-500	White and pink sandstone with thin layers of shale; surface cavernous.
EOCENE	Poison Canyon fm. Possibly including beds equivalent to Raton formation.		2000-3500	Arkosic sandstone and fine conglomerate, with thin beds of yellow clay; lower beds weather pale yellow.
UPPER CRETACEOUS	Vermejo fm.		0-450	Dark shale, light gray friable sandstone and coal in alternate layers.
UPPER CRETACEOUS	Trinidad ss.		160-225	Massive sandstone, shaly in lower part.
UPPER CRETACEOUS	Pierre shale		1800-2000	Yellowish gray to dark gray shales, with zone of impure limestone concretions.
UPPER CRETACEOUS	Apishapa sh.	OF NIOBRARA AGE	450-500	Bluish gray shales at base, grading upward through papery shales to sandy shales.
UPPER CRETACEOUS	Timpas ls.	OF NIOBRARA AGE	180-200	Grayish white ls. and calcareous sh.
UPPER CRETACEOUS	Carlile sh.	OF BENTON AGE	170-180	Dark gray shale, capped by yellowish ss.
UPPER CRETACEOUS	Greenhorn ls.	OF BENTON AGE	30-40	Thin-bedded dove colored limestone.
UPPER CRETACEOUS	Graneros sh.	OF BENTON AGE	200-310	Gray to black sh. with concretions near base.
LOWER CRET.	Dakota ss.		350-400	Dense, fine grained sandstone. Coarse, yellowish gray ss., overlain by thin bed of gray shale.
LOWER CRET.	Purgatoire fm.			
UP. JURASSIC	Morrison fm.		100-240	White sandstone, pink and green shales and hard fine grained limestone.

STRATIGRAPHIC COLUMN OF MESOZOIC AND TERTIARY FORMATIONS OF HUERFANO PARK AND VICINITY

Dawson arkose and the Denver formation. Both the Poison Canyon and the Dawson are arkosic facies of sedimentation. Although the Poison Canyon was considered by Lee to overlie the Raton formation, Knowlton regarded the flora of the two formations as indistinguishable. Lee[32] pointed out, also, that it is not always possible to distinguish the rocks of the two, and that, although the contact between the two formations has generally been regarded as unconformable, it is difficult to trace. It may be that the formations are nearly equivalent in age and that their stratigraphic relations are complex and interfingering, as Richardson has shown to be the true relation between the Denver formation and the Dawson arkose.[33]

The coarse arkosic nature of the soft grits comprising the Poison Canyon formation reflects the earlier, though not strictly the beginning, stages of the Laramide revolution, which consisted in initial uplifting of the granitic massifs of the Paleozoic landmass to the southwest, with the accompaniment of local erosion of the Trinidad and Vermejo formations. This is represented in the Huerfano Park section by the hiatus between the Pierre shale and the Poison Canyon formation.

The Cuchara formation overlies the Poison Canyon beds, with marked unconformity at the northern end of Huerfano Park and entirely overlaps the formation locally. As the Cuchara formation is, in turn, overlapped along the eastern side of the park by the Huerfano beds, it is assigned questionably to the lower Eocene (Wasatch) epoch.

Exploration by Osborn and Wortman in 1896 and by Granger and Olsen in 1918 has established,[34] on the basis of vertebrate remains, the fact that the upper third of the Huerfano formation is to be correlated with the upper Wind River and the lower Bridger beds of Wyoming. According to Osborn,[35] the deposition of the formation apparently began near the end of Wasatch time and extended into early Bridger time.

Physiographic and Climatic Conditions of Bridger Time.—Inasmuch as physiographic conditions during deposition of the Huerfano beds have an important bearing on the structural history of the Sangre de Cristo region, it is of interest to consider briefly Osborn's conclusions applying generally to the life environment of the Bridger fauna. Although Osborn's analysis applies strictly to the Bridger Basin of Wyoming, the fact that both the Huerfano fauna and the nature of the sedimentation are, in

[32] W. T. Lee and F. H. Knowlton: *Geology and paleontology of the Raton Mesa and other regions in Colorado and New Mexico*, U. S. Geol. Surv., Prof. Pap. 101 (1917) p. 61.

[33] G. B. Richardson: *Description of the Castle Rock quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 198 (1915) p. 7.

F. H. Knowlton: *op. cit.*, p. 4.

[34] H. F. Osborn: *The titanotheres of ancient Wyoming, Dakota, and Nebraska*, U. S. Geol. Surv., Mon. 55, vol. 1 (1929) p. 74-75.

[35] *Op. cit.*, p. 74.

some respects, similar to those of the Bridger basin makes a comparison justifiable. The comparative conclusions are, of course, subject to certain qualifications imposed by distance and by local differences in critical factors.

Osborn [36] states that the deposits of the Bridger Basin were largely the result of fluviatile and flood-plain sedimentation, with swamps and fresh-water bays bordering the lowlands. The Huerfano beds comprise sandy clays, soft sandstones, and shales, incised here and there by channel sandstones, and, hence, were doubtless deposited under conditions similar to those of Wyoming. Though these Huerfano deposits were referred to as "lake beds" in the older literature, it is noteworthy that the fresh-water limestones and varved deposits commonly associated with the ponding action of mid-Tertiary volcanism are inconspicuous or absent. The clay deposits, as well as the presence of reptilian remains, attest, however, to the existence of local swampy conditions. With regard to the climatic implications of the Bridger fauna, Osborn further states:

"Crocodiles were numerous and diversified. The reptiles suggest that the climate was Floridan, or south temperate, and we may picture a partly open, partly forested country, somewhat similar to the existing bayou region of the Mississippi delta of Louisiana. . . . The foot structure of the Bridger quadrupeds gives less certain evidence of an open plains country, favorable to the cursorial types, than that of the Wasatch (lower Eocene) quadrupeds of the same region." [37]

However, as the Huerfano fauna is less fully described, and perhaps less diversified, these conclusions are probably not directly and fully applicable in the present state of knowledge.

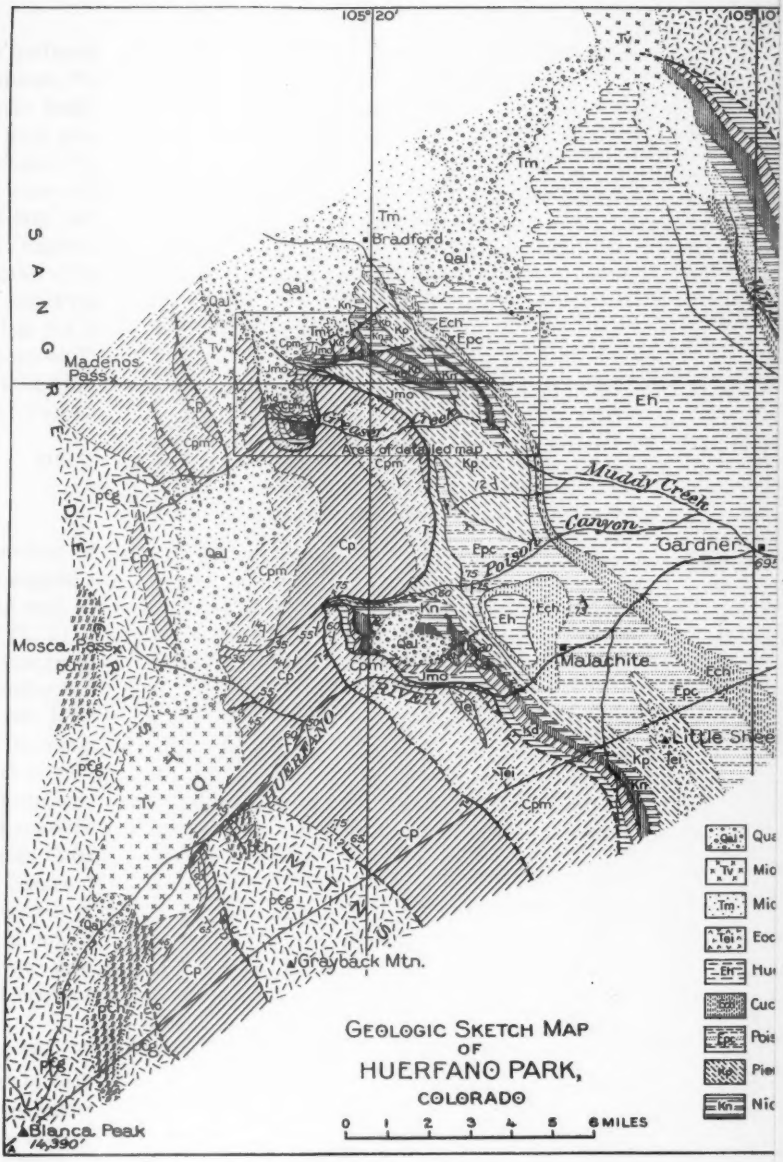
Even with these reservations, however, it is believed that the faunal and climatic evidence tends to corroborate the evidence of sedimentation—that, although mountainous tracts may have existed, they were either of moderate height or were situated much farther from the basin than are the existing massifs (Pls. 1, 4). In later Huerfano time the increased proportion of conglomeratic debris in the streams, as shown by the upper part of the formation near the Spanish Peaks,[38] seems to indicate the beginnings of differential uplift of the nearby pre-Cambrian massifs. This movement culminated with the elevation of the mountains to near their present heights of 4,000 to 6,000 feet above the top of the Huerfano formation. This post- or late-Bridger uplift of the pre-Cambrian massifs and its effect upon the structure of the mountains will be described in subsequent pages.

---

[36] *Op. cit.*, p. 80, 81.

[37] *Op. cit.* p. 80, 81.

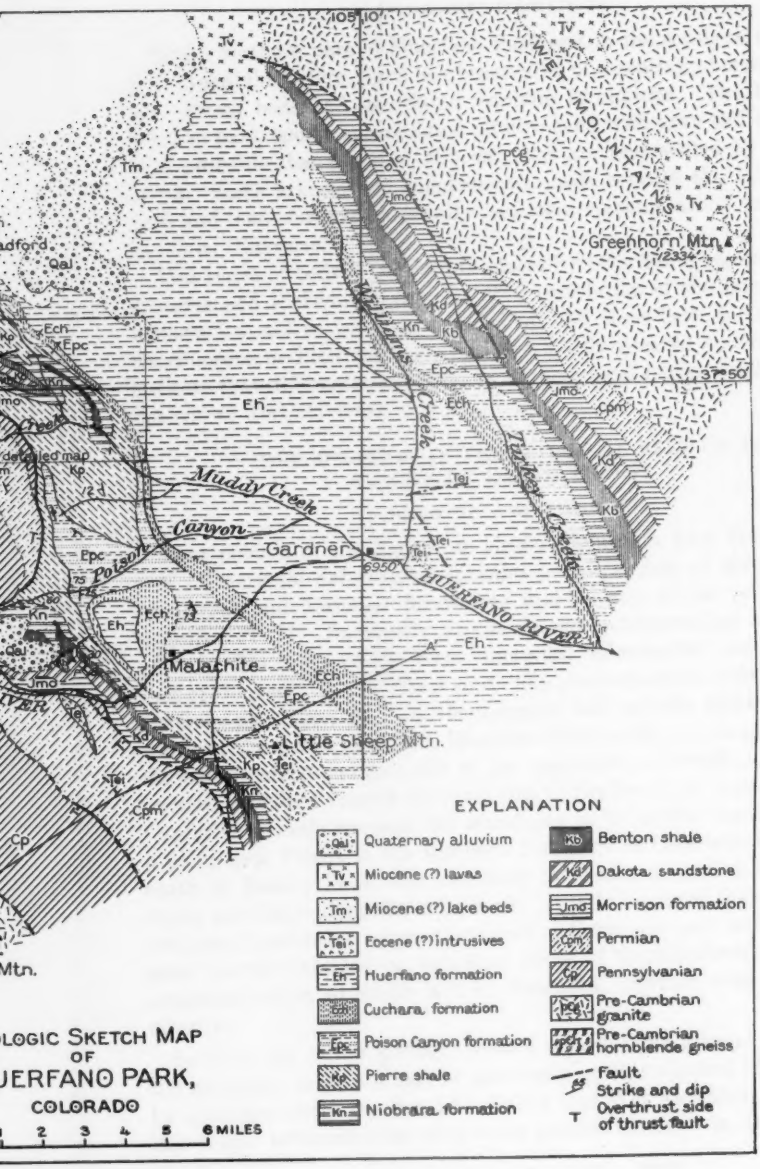
[38] R. C. Hills: *Description of the Walsenburg quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 68 (1900) p. 2.



GEOLOGIC SKETCH MAP OF HUERFANO PARK, CO

GEOLOGIC SKETC
OF
HUERFANO
COLORAD
0 1 2 3

GEOLOGIC SKETCH MA



OLOGIC SKETCH MAP
OF
UERFANO PARK,
COLORADO

SKETCH MAP OF HUERFANO PARK, COLORADO

POST-EOCENE FORMATIONS

In the northern part of Huerfano Park and in the southern end of the Wet Mountain Valley, there is a broad area of essentially horizontal lake and stream beds and associated volcanic formations. The beds comprise, in part, poorly consolidated pinkish sand with gray tuff at the top. It is probable that parts of Miocene time are represented by these deposits. Some arkosic and conglomeratic beds beneath the Miocene (?) tuffs and lake beds are essentially free from volcanic debris and may be as old as Oligocene.

This series of beds is overlain in places by lava flows, breccias, and by Pliocene and Pleistocene gravels. The correlation of the later Tertiary deposits has not as yet been established. All these beds, including even the Pliocene or Pleistocene gravels, are locally fissured and faulted and indicate that differential movements of the crustal blocks continued after the end of Tertiary time.

DEFORMATION IN THE MARGINAL BELT OF THE SANGRE DE CRISTO THRUST

DEFINITION AND SUMMARY

That portion of the Sangre de Cristo Mountains here defined as the belt of marginal deformation includes only the zone of deformed Carboniferous and Mesozoic formations that lie east of the pre-Cambrian massifs (Pl. 1). As closely as the present interpretation of structure permits, this usage of the term "marginal deformation" corresponds to that of Bucher.[39] From Orient south to a position about midway between Crestone and Huerfano Park the marginal belt includes almost the whole width of the range but, from Huerfano Park south, it comprises chiefly the eastern slope and foothills of the mountains, although, in places, its western boundary crosses the crest line to the San Luis slope. The pre-Cambrian massifs compose the remaining width of the mountains, from a point near Orient to the southern boundary of Colorado, except that, south of Blanca Peak, this basement is locally concealed by Tertiary lavas, especially along the west flank. It is to be noted that the massifs comprise, in addition to pre-Cambrian metamorphic and intrusive rocks, some narrow infolded or infaulted slices of Carboniferous strata, the structural origin of which will be discussed, together with that of the massifs.

Some of the earlier geologic writers regarded the general structure of the mountain belt as a narrow and compressed anticlinal fold, bounded by synclines under the San Luis Valley and Wet Mountain Valley; but this simple interpretation must be so greatly modified in its application

[39] W. H. Bucher: *The deformation of the Earth's Crust* (1933) p. 149-161.

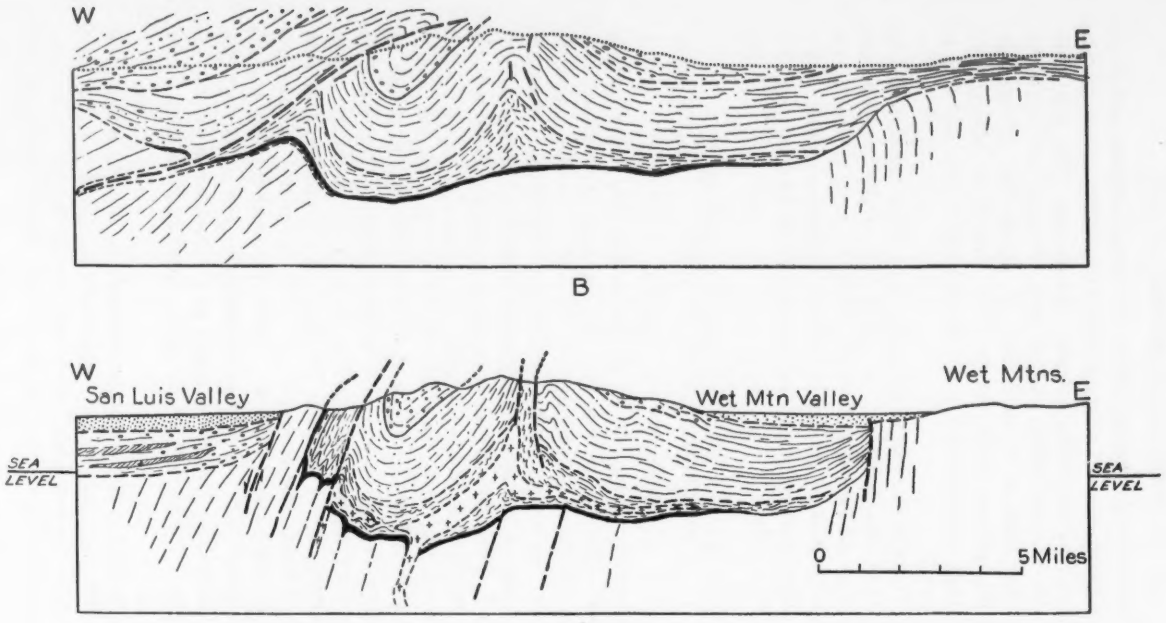


FIGURE 1.—*Generalized and diagrammatic sections across the north-central part of the Sangre de Cristo Mountains*

Section A constructed chiefly from a traverse across the range near Crestone. Symbols are generalized: solid black pattern (base) represents lower Paleozoic sandstones and limestones; overlying crinkled line pattern, lower Pennsylvanian beds involved in zones of shearing and plastic deformation; lighter line and conglomerate pattern, higher Pennsylvanian and Permian formations; the filling of the San Luis Valley depression is late Tertiary and Recent alluvium, and older lavas and gravels; intrusive bodies are entirely hypothetical but are at positions corresponding to similar bodies exposed in other parts of the range. Section B is a diagrammatic representation of the same section, based upon the hypothesis of an earlier phase of lower-angle overthrusts (pre-Bridger), which are presumed to be responsible for the greater part of the shallower tangential deformation in the marginal belt.

to different parts of the range that it is not adequate. In the following description the marginal belt, as delimited, will first be considered independently of the massifs, for it is believed that certain phases of deformation have acted differently within the two tectonic units. Those portions of the Sangre de Cristo Mountains between Orient and the Arkansas River and between the south end of Huerfano Park and the Colorado-New Mexico line will be left out of the descriptions referring to different stages of marginal deformation, chiefly because they have received little study. It is known, however, that the northern part of the range, between Orient and the Arkansas River, lies outside the immediate influence of the principal overthrust, which diverges from the mountain range near Orient and crosses the head of the San Luis Valley, to its west side. Moreover, the structure of this northernmost segment of the range is transitional between the Sangre de Cristo belt itself and the faulted synclinal folds that connect with the structural province of South Park, west of the Colorado Front Range (Pl. 1).

Between Orient and Huerfano Park, two principal kinds of structure—that is, complex folds and overthrusts—dominate the marginal belt. The folds extend S. 35° E., from a point in the vicinity of Orient to a point beyond Crestone, and consist of a major central anticline, bounded on the west by a much compressed, overturned, and faulted syncline, and on the east by a more open syncline, which appears to be continuous with, or merges into, the major syncline that extends north to the Arkansas River (Pl. 1). The major central anticline is complicated by the presence of a metamorphosed core and by several small intrusive igneous bodies along, or near, its axis. The overthrusts are best preserved in the Huerfano Park province (Pl. 4). They comprise several separate thrust plates, imbricated or lapped one upon another and thrown into steeply inclined positions by the forces that caused marginal deformation. This structure is complicated by local recumbent folding of the frontal thrust plate. These kinds of structure will be briefly illustrated by descriptions of two typical cross-sections of the range.

NORTH-CENTRAL PORTION OF THE RANGE

The older anticlinal interpretation of the structure seems to find some support in the 9-mile cross-section of the range between Orient and Crestone. Approaching the mountains from the San Luis Valley, in the vicinity of Crestone, one first encounters the pre-Cambrian schists and dioritic rocks, which rise abruptly from the valley along a linear mountain front. The strip of pre-Cambrian rocks comprising the massif is less than a mile wide at this locality and is bounded on the east by overturned Paleozoic formations, which dip about 75 degrees westward. The Harding sandstone, of Ordovician age, which forms part of the massif,

lies against the granitic rocks in normal contact and has not been sheared from its base. Beyond it, the Fremont limestone, the Chaffee formation, and the Leadville limestone are exposed in normal stratigraphic succession, but likewise overturned. The limestones, shales, and sandstones of these younger formations become progressively more deformed by plastic flow and shearing toward the east. The Leadville limestone is schistose and flaky, and its normal thickness seems to have been greatly reduced. Numerous bedding faults in the Carboniferous strata above the limestone make it uncertain whether or not the Pennsylvanian section is represented in its entirety. To the north of this position, on Crestone Creek and higher on the flanks of the range, these lower Paleozoic and Pennsylvanian beds have an overturned dip of 45 or 50 degrees westerly, indicating that locally the basement rocks have welled up and overridden the relatively depressed block of sedimentary beds. These structures, incidental to upthrusting and overthrusting along the edge of the pre-Cambrian massif, were probably formed at a later time than structures that characterize the marginal belt itself, which forms the remainder of the cross-section of the range.

The synclinal limb of highly compressed and deformed Pennsylvanian beds extends from one mile to one and a half miles east of the basement, where the massive Permian conglomerates stand vertically in contact with the Pennsylvanian shales. The Permian conglomeratic beds form a large asymmetrical syncline, which extends nearly to the crest of the range, one and a half to two miles farther east (Pl. 5). The west limb of this syncline, as already mentioned, is vertical or slightly overturned, whereas the east limb dips from 20 to 40 degrees west. East of this synclinal body of conglomeratic beds, Pennsylvanian marine beds with steep westerly dips of 50 or 55 degrees are again exposed, thus forming the east limb of the syncline. The dips in the east limb increase eastward until, at a position a mile or less east of the crest of the range, the beds stand vertical and form the core of the central anticline. From the anticlinal core east to the edge of the Wet Mountain Valley the flank of the range is formed of Pennsylvanian and overlying Permian beds in normal, and apparently undisturbed, succession and with decreasing easterly dips. They form the western limb of a syncline whose eastern limb is concealed beneath the alluvial deposits of the Wet Mountain Valley. This syncline obviously represents less deformation than that flanking the west side of the range.

With regard to the character of deformation, the central anticline of the range is of special significance. Both in the section just described and in the area east of Orient, the Pennsylvanian shales and sandstones become progressively more metamorphosed as the core of the anticline

PERMIAN SYNCLINE

View looking south at the head of Willow Creek Park, east of Crestone, central western slope of the Sangre de Cristo Range. Prominent ledge (above men and horses, right center) is composed of Permian conglomerate red beds and dips southwest 70 to 80 degrees overturned. It is in the western slightly overturned limb of the syncline, the axis of which approximately coincides with the U-shaped valley in the center. Underlying the park (low foreground) are the highly compressed Permian and Pennsylvanian beds that form the west flank of the marginal belt. (Photograph by C. E. Siebenthal, United States Geological Survey)

FIGURE 1. VIEW ACROSS UPPER PART OF POISON CANYON
An overthrust plate of the Pennsylvanian beds forms the hill in the background. Niobrara limestone, dipping north beneath the thrust, forms low ridges in the middle foreground. The valley between is underlain by Pierre shale upon which the thrust plate lies.

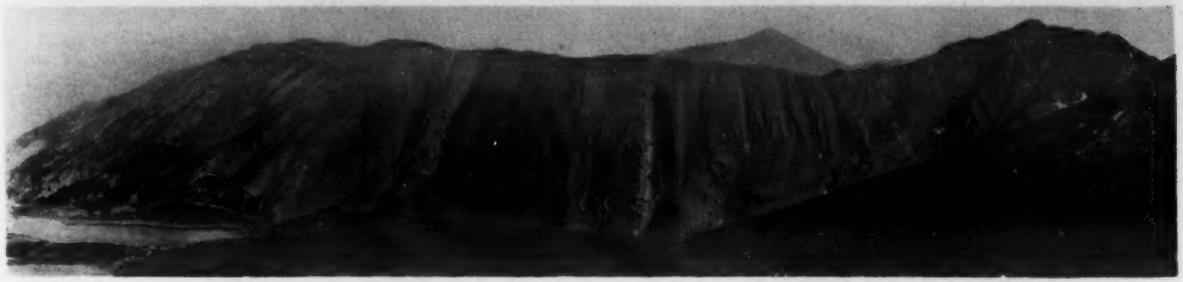


FIGURE 2. LOWER PENNSYLVANIAN BEDS IN THE CORE OF THE RANGE
View south to southeast from the divide east of Crestone, showing the steep dips. Dark band of essentially vertical beds (center) constitutes the metamorphosed schistose core of the fold.

VIEWS OF STRUCTURAL FEATURES

is approached. The core itself is composed of dark-colored micaceous sandstones and shales that are contorted and locally converted to schists. The width of this central core is comparatively narrow, probably not exceeding half a mile, but the surrounding zone of steeply dipping or nearly vertical Pennsylvanian beds has a width ranging from one to two miles (Pl. 6, fig. 2). As the marine shales, limestones, and sandstones of the core are known to exist only in the lower part of the Pennsylvanian series, it appears that these beds have been forced upward, and perhaps in part laterally, by combined folding and plastic flow from near the base of the Carboniferous section. So far as can be recognized, no rocks in this core belong to the pre-Cambrian basement complex. Both at Crestone and at Orient, the plastic flowage has been sufficient to disrupt the central anticline by zones of shearing and faulting, so that, in general, there exists a discordance between the core and the limbs of the fold, such as is common to injective or diapiric folding. Locally, drag folds are exposed in the beds of one or both limbs, but have not been given the study needed to warrant interpretation. Plate 6, figure 2, shows an apparent large drag fold on the east limb of the central metamorphosed anticline east of Crestone. This fold was not viewed except from a distance, but does not appear to agree with the expected direction of drag resulting from diapiric folding. Numerous small drag folds associated with the much-contorted core of the fold afford opportunity for a more conclusive study. In the Crestone section the stratigraphic relations indicate that the western limb of the anticline has been relatively uplifted, but whether or not igneous bodies have contributed to this structure is not apparent. Farther north, however, in the area east of Orient, narrow, elongate or elliptical intrusive masses, from half a mile to a mile and a half in width, are exposed in the west limb of this anticline. These intrusives are of quartz dioritic to granodioritic composition, with local bodies of coarse porphyritic granite. It may well be supposed that this narrow core of metamorphosed rocks, together with the associated coarse granitic rock, led to the earlier views that the range was a sharp anticline with a pre-Cambrian core. The principal secondary mineral of the dark-colored schists is biotite. Locally, other contact minerals are developed, and the bodies of igneous rock doubtless influenced the general metamorphism of the core as well. From views afforded on the high central peaks, it appears that this partly metamorphosed core extends continuously, from the vicinity of Orient southward along the crest of the range, for 25 miles, and that at least three small intrusive bodies flank it on the west side, as shown in Plate 1.

In Figure 1A, an attempt has been made to construct a typical cross-section of the range, such as that described near Crestone, showing the

present structure extended hypothetically in depth. Figure 1B represents a supposed early stage of overthrusting prior to the upthrust of the pre-Cambrian rocks and the downfaulting of the San Luis Valley, and is discussed further in subsequent pages. The central anticline is the unique feature of the marginal belt and dominates the structure in the north-central part of the range. The details of deformation in the core may be compared to those in diapiric or injective folds, such as have received much attention in the literature of salt domes and certain mountain ranges in the oil-producing regions of Europe. Recently, Wegmann [40] has discussed the mechanics of deformation in diapiric structures and has applied it to intrusive bodies and structures in the pre-Cambrian basement ranges of Finland. The principles of the structure appear to have application outside of salt-bearing terrains. According to Wegmann, the fundamental factors controlling this type of structure, and upon which there is general agreement, are the existence of a bed, or a series of beds, of highly mobile rocks that lie deeply buried beneath less mobile and generally specifically heavier rocks. Within the mobile layer the material moves by plastic flowage toward an egress or zone of less pressure and rises toward the surface, resulting finally in shallower injective structures. There is no general agreement as to whether earlier tectonic movements initiate the arching or whether the differences in plasticity and thickness of the original formations are the chief causes of localization.

In the Sangre de Cristo Range, these fundamental conditions are recognizable in the existence of the relatively incompetent series of shaly beds in the lower Pennsylvanian, which are overlain by a great thickness of less mobile and heavier conglomeratic beds. The localization of this linear belt of plastic deformation in the range seems to be dependent, in part, upon a period of thrusting, for its greatest development lies just east of the arc-shaped bend in the Sangre de Cristo thrust zone, whereas it disappears from the range in the region just north of Orient, not far from the point where the thrust zone leaves the range and crosses the San Luis Valley.

However, particular significance must be attached to the fact that the existence of true schists in the Paleozoic formations is confined strictly to those zones of deformation where underlying or near-by intrusive bodies may have exerted contributory influence. Near other zones of deformation in the range, there is, commonly, moderate to intense shearing but little recrystallization, whereas the intricate foliation and plastic deformation (such as that seen locally in the central core, near intrusive

[40] C. E. Wegmann: *Ueber Diapirismus (besonders im Grundgebirge,)* Finland Geol. Komm., Bull. no. 92 (1930) p. 58-76.

bodies) is entirely wanting. The conclusion seems to be warranted that mineralizers saturated the rocks surrounding the igneous bodies, and induced sufficient melting and softening to enhance the plastic deformation of the rocks. It is also noteworthy that the lines of schistosity commonly parallel the bedding planes of the sedimentary rocks, though, locally, these planes are strongly imbricated in zones of more intense movement.

This problem, relating to the mode of formation of the Carboniferous schists, is one that deserves more intensive study. Both the central core of the range and an area south of Orient, west of the central anticline, are well adapted to studies of gradations between the normal sedimentary rocks and the most advanced stages of metamorphism.

HUERFANO PARK REGION

General Structure of Deformed Belt.—The second principal kind of structure characterizing the marginal belt of the range is that found in the zone of overthrusting at Huerfano Park. Here, the east flank of the Sangre de Cristo Range is of special interest, not only because of the existence of imbricated thrust plates, but also because here the latest Cretaceous and Tertiary formations are involved in the deformation. In other parts of the range, only Paleozoic or Mesozoic formations are involved. The structure zones common both to the park and to parts of the range farther north thus permit the determination of approximate ages of some of the principal events of deformation. Although the region between Huerfano Park and Crestone has not been adequately explored, there can be little doubt that the compressed metamorphosed folds change southward into imbricate structures that characterize the marginal belt west of the park. There is one feature common to both kinds of structure—the overlapping thrust plates, like the close folds, contain no formations stratigraphically below the lower Pennsylvanian shales.

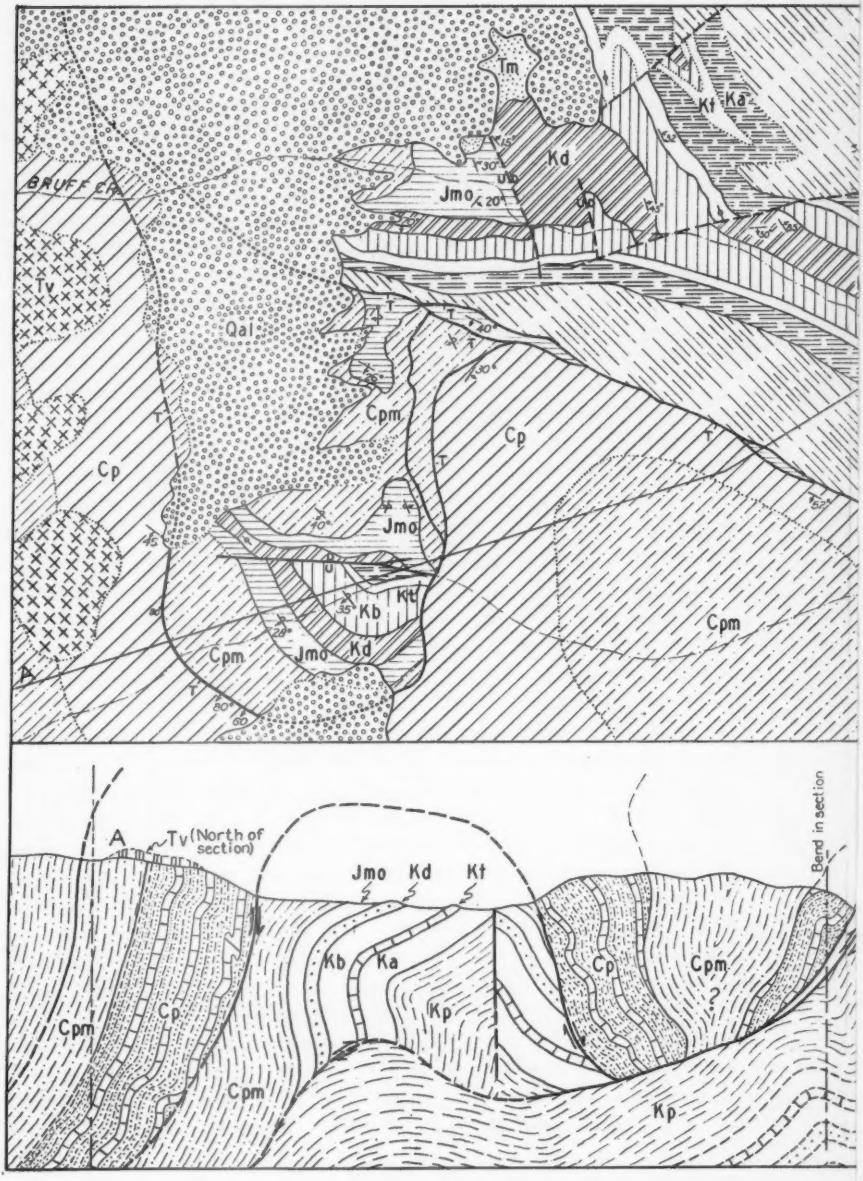
An illustration of the frontal thrusts of the area is shown in Plate 7, which shows part of a lobate salient of the main thrust belt, near Bruff Creek, a tributary of the Huerfano River drainage. The sharp folds in the Cretaceous formations have been produced by the overriding movement of late Carboniferous strata, which form the upper plate or sheet of the overthrust. The composite sole of the overthrust plate at this locality consists of two thrust surfaces, which at some places are separated and well defined, and at others the two surfaces so nearly coincide that their separate movements cannot be distinguished. North of Greaser Creek, where the two are separated, the older thrust surface lies above the younger and has been sharply folded. This folding involves beds beneath the thrust surface, which is evidenced by the inlier of Cretaceous formations. Mesozoic and Permian beds of this inlier have been de-

formed into an inverted pseudo-dome (Pl. 7, Section AA), having the Timpas limestone, the Apishapa shale, and the Dakota sandstone exposed on the crest, with Morrison and upper Permian beds on the flanks. This inversion of the strata beneath the upper fault surface is interpreted to be the result of an easterly directed salient of the thrust, which overturned the beds into a recumbent fold that finally sheared above its under limb. With continuation of thrusting along a deeper fault, the front of the thrust plate became incompetent and was locally folded. Later movement finally produced the lower thrust surface, which broke in advance of, and near, the older folded thrust. The two thrusts might be the result of bifurcation of a single major thrust at depth. The stratigraphic throw of these faults is roughly from 13,000 to 15,000 feet, assuming a minimum thickness of 8,000 feet for the upper part of the Carboniferous formations. The shortening produced is difficult to estimate, but is probably not less than 5 or 6 miles for this one structure. A typical view of the frontal thrust plate on the north side of Poison Canyon, a few miles south of the Bruff Creek area, is shown in Plate 6, figure 1.

It is significant that the floor of the frontal thrust is made up of the plastic Pierre shale.

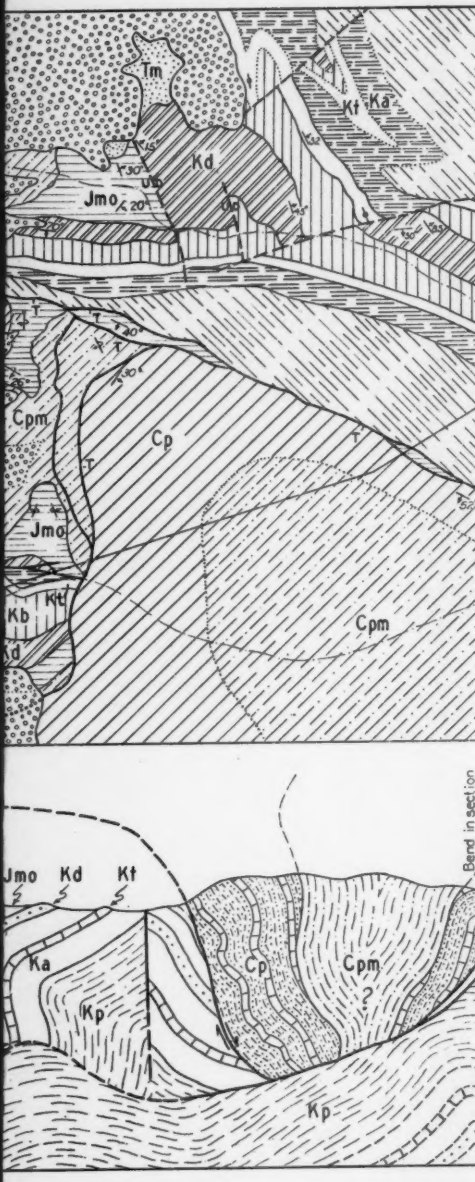
Between this zone of frontal deformation and the crest of the range, 4 or 5 miles to the west, parts of the Pennsylvanian and Permian beds are repeated several times by steep thrust faults. In each thrust slice the beds dip at angles ranging from 60 degrees to vertical, and lie with their tops to the west. In the first thrust plate west of the frontal faults, there are 5,000 to 6,000 feet of Pennsylvanian and Permian beds, consisting of both marine and terrestrial deposits. Toward the west, the thickness of marine beds becomes less in each successive slice, and the terrestrial deposits change rapidly in texture to coarser facies. On the west side of the range, in the valley of Madenos Creek, a body of coarse conglomerate, representing alluvial fan and talus deposits, appears to rest directly on pre-Cambrian granite. These conglomerates are believed to belong to the upper part of the Permian conglomeratic red beds. The amount of movement between the imbricated slices cannot be estimated without assumption as to the total depth of folding, although the change in facies from one slice to the next and the character of the marine deposits suggest that the basin in which these formations were deposited has been much foreshortened.

In a later section on deformation within the pre-Cambrian massif that lies west of this belt of imbrication (p. 969 and Fig. 2), evidence is presented to show that late uplift and lateral expansion of the massif have tended to tilt the upper portion of the plates eastward. In view of their present vertical to steep westerly dip, it is possible that, during the cul-



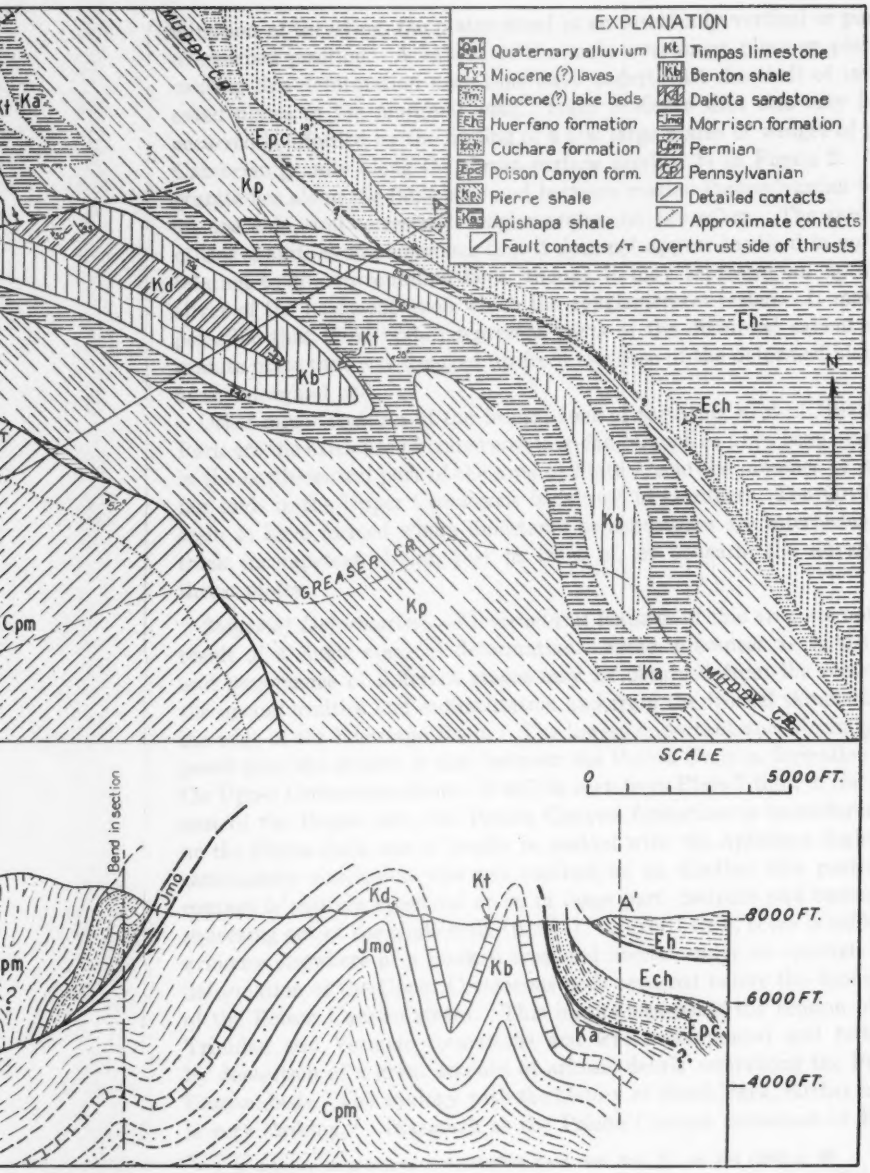
PLANE-TABLE MAP AND STRUCTURE SECTION OF A LOBA

Bruff Creek area, Huerfano Pa



-TABLE MAP AND STRUCTURE SECTION OF A L

Bruff Creek area, Huerfan



TION OF A LOBATE SALIENT OF THE MAIN THRUST BELT

rea, Huerfano Park, Colorado.

mination of thrusting, the plates stood in an essentially vertical or partly overturned position. A salient of the main overthrust plate or plunger causing this deformation may thus have underthrust the shell of imbricated plates in the marginal zone. Some evidence that this may have occurred is afforded by the finding of a few large plates or wedges of pre-Cambrian granite along the thrust surface marked f1 in Figure 2. The granite lies along the fault zone and between marine Pennsylvanian beds, on one side, and terrestrial conglomerates, on the other. The position of this granite is so anomalous that it does not appear to stand in normal relation to the surrounding beds, though it may represent an upthrust sliver of the basement upon which the conglomerates rested. It may be possible that these bodies represent parts of the disrupted salient of a granite plate that was dragged up from the base of the deformed blanket of the marginal belt.

Other minor features of the frontal zone of thrusting west of Huerfano Park are the existence of transverse tear-faults, such as have been mapped in the north-central part of the area of Plate 7, and the presence of small fan folds in the Upper Cretaceous beds west of Gardner. The fan folds seem to have formed where limestones or sandstones are interbedded in thick incompetent shales, just in front of, or beneath, the advancing thrust plates.

Stages of Deformation.—All structural features of this area are not the result of a single stage of deformation. Some inferences as to the age and importance of different stages may be obtained from the record of the unconformities and sedimentation and from the frontal thrusts along the edge of the Huerfano Basin. The lowest important unconformity exposed near the thrusts is that between the Poison Canyon formation and the Upper Cretaceous shales. It will be seen from Plate 7 that, in the zone east of the thrust lobe, the Poison Canyon formation is unconformable on the Pierre shale and is locally in contact with the Apishapa shale. A satisfactory conclusion was not reached as to whether this particular contact is entirely erosional or is, in large part, tectonic and caused by squeezing out of the shale beds (p. 961). Nevertheless, there is sufficient evidence elsewhere of a marked erosional unconformity to conclude that deformation of the Upper Cretaceous beds occurred before the deposition of the Poison Canyon strata. This hiatus represents the erosion of the Trinidad and Vermejo formations and was accompanied and followed by deposition of a great amount of arkosic debris comprising the Poison Canyon beds. The analogy with the history of South Park, farther north, is very striking.[41] Inasmuch as the Poison Canyon formation of Huer-

[41] J. H. Johnson: *Stratigraphy of South Park*, Geol. Soc. Am., Pr. for 1934 (1935) p. 86.
C. H. Behre, Jr., I. T. Schwade, and R. M. Dreyer: *Bedrock geology of northern South Park*, Geol. Soc. Am., Pr. for 1934 (1935) p. 67.

fano Park contains a comparatively small proportion of debris derived from Carboniferous beds and a large amount of granitic debris, this early deformation evidently resulted in uplift and erosion of old Paleozoic landmasses upon which the Carboniferous formations were thin or absent. The broadest of these landmasses lay west and southwest of Huerfano Park and included the Paleozoic San Luis Range and the broad pre-Cambrian massif backing it, now largely covered by the San Juan volcanic rocks. The Front Range of Colorado, to the north, and possibly parts of the Wet Mountains were also exposed to erosion at this time. The culmination of this earlier stage of deformation would seem to have occurred sometime after Fox Hills and Laramie times but before Fort Union time, as the plants of the Poison Canyon formation correspond more closely to the Dawson and the Raton types than to the late Fort Union flora.

The degree and extent of deformation in Paleozoic and Mesozoic formations during this stage may be partly inferred in like manner from the sedimentary record. In Huerfano Park the overthrust plates of Carboniferous strata now lie upon the upper part of Pierre shale, a position to which they could not have been thrust previous to the epoch of Poison Canyon sedimentation, else the resulting bulge would have caused the rapid erosion of the softer Mesozoic beds and the incorporation of much debris from the Carboniferous thrust plates in the Poison Canyon and Raton formations. Though the full thickness of the Poison Canyon formation is perhaps not preserved, and perhaps none of the Raton is present in this immediate area, neither are these formations elsewhere of the nature expected under such conditions. Thus, even though folding and tilting must have been initiated, as shown by the great unconformity at the base of the Poison Canyon, overthrusting of the Carboniferous onto the Cretaceous formations could hardly have developed beyond incipient stages.

The Cuchara formation overlies the Poison Canyon formation with marked angular and erosional unconformity, and locally overlaps onto the Pierre shale, so that the Poison Canyon formation in the area shown in Plate 7 is greatly reduced from its normal thickness a few miles to the south. The Cuchara rests directly on the Pierre shale at one place in the eastern part of the area of Plate 7, and also just south of Bradford, as shown in Plate 4. The Cuchara, compared with the other formations, is characterized by the more heterogeneous composition of its debris. The beds contain not only a greater variety of pre-Cambrian fragments, but also some conglomeratic volcanic material. In the conglomerates are fragments of Paleozoic sedimentary rocks that are indicative of a comparatively close source of debris. In turn, the Huerfano formation lies

with angular unconformity upon the Cuchara, where the latter formation has been sharply upturned by salients of the thrust plates, as shown in Plate 7. Debris composing the Huerfano beds has been clearly derived by the erosion of land areas that consisted chiefly of Permian beds. The resemblance between the reddish gritty shales of the lower part of the Huerfano formation and the red Permian shales is so close that it is difficult to distinguish lithologically certain parts of the two formations. The debris deposited in the lower Huerfano beds differs chiefly from that of the Permian by a smaller content of mica. Hills states, with regard to the Huerfano beds: "It was the erosion of this conglomerate [upper Carboniferous Sangre de Cristo formation of Hills] that furnished the bulk of the material composing the Huerfano beds of the Spanish Peaks area."[42] In the Spanish Peaks quadrangle, the Huerfano beds contain much more conglomerate in the upper part than they do in the Huerfano Park region.

Both the Huerfano and the Cuchara formations were affected by the overthrusting, but the differences between them in the degree of unconformity and involvement in open folding are much less than between the Cuchara and the Poison Canyon beds. This difference is indicative of declining intensity of the overthrusting in the later epochs. Furthermore, the Huerfano beds in the center of the basin lie nearly horizontal and with little apparent crumpling other than that close to the front of the thrust fault. Thus, deformation of the underlying Mesozoic and Paleozoic strata must have been of minor extent during later Huerfano time (Wind River to early Bridger); otherwise, the crumpling of more rigid beds at depth beneath the edge of the basin would have resulted in noticeable upbulging marginal to the thrust plates, such as shown by the Poison Canyon and older Cretaceous formations.

From these facts, it may be inferred that deformation continued active through Cuchara into Huerfano time, and that overthrusting of the Carboniferous strata must have reached an advanced stage between the deposition of the Poison Canyon formation and that of the Cuchara formation, but continued with lessening intensity into Huerfano time. The stages resulting in the overriding and imbrication of thrust plates would be of post-Fort Union and pre-Wind River age, perhaps reaching their greatest development in early Wasatch. Lack of knowledge of the exact age of the Cuchara formation prevents a closer dating of the earlier stages.

These conclusions, based chiefly upon the sedimentary record and the unconformities between the formations, are seemingly only partially substantiated by the degree to which the Poison Canyon beds are involved

[42] R. C. Hills: *Description of the Spanish Peaks quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 71 (1901) p. 2.

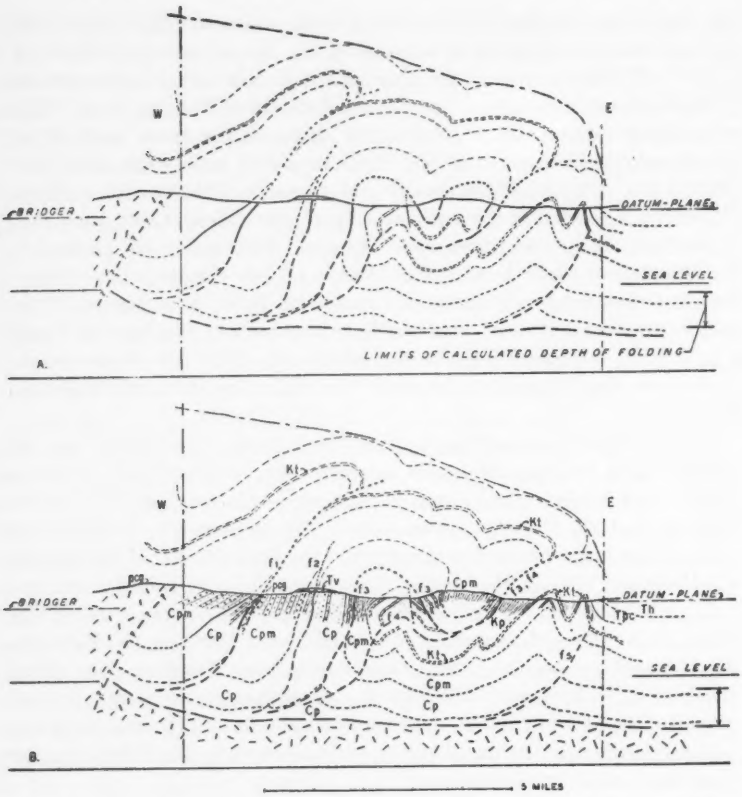


FIGURE 2.—*Geologic section from the Bruff Creek area west to the pre-Cambrian massif south of Madenos Pass*

Shows reconstruction of faulted and folded beds above and below the datum plane representing approximate base level in Bridger time. Sections A and B illustrate two different interpretations of the frontal folded thrust plate.

in deformation along the west side of Huerfano Park. Here, a salient of the thrust plates, partially illustrated in Plate 7, has caused different degrees of deformation in formations marginal to its front. The Poison Canyon formation noticeably is not involved in the more highly contorted deformations exhibited by the Timpas limestone and other formations below the Pierre shale. In general elsewhere, the formation is involved in the more open folds of the older formations and locally is overturned near the thrust faults. This difference may perhaps be accounted for largely by several factors unrelated to any earlier or additional stages of deforma-

tion to which the older formations have been subjected. The full extent of deformation in the Poison Canyon beds may not be entirely exposed in the area shown in Plate 7, inasmuch as the formation is concealed to the east by younger beds and has been entirely removed locally near the thrusts by pre-Cuchara erosion. Of greater importance, however, is the fact that the Pierre shale, nearly 2,000 feet in thickness, tended, as a result of plastic flowage in the formation, to disperse the directed forces of the advancing thrust plates and to buffer the thrust plates by a plastic body that bulged upward and elevated the more rigid Poison Canyon formation. Hence, the Poison Canyon and other comparatively competent Cretaceous formations above the Pierre shale were rarely subjected to the direct impingement of the thrust plates upon them. The fact that the Cuchara, Poison Canyon, Raton, Trinidad, and Vermejo formations have not been overridden by the Carboniferous rocks anywhere along the front of the Sangre de Cristo Mountains is perhaps due to this buffer action of the underlying thick section of shale beds. The still lower shales of the Permian red beds are possibly of some influence in deeper zones, inasmuch as the Dakota sandstone is the thickest of the underlying more rigid formations, and there are only a few sandstone beds in the Morrison. At least a thousand feet of soft Permian red shales are present beneath the Morrison formation in the Huerfano Park region. Thus, the fact that the Pierre shale is thinnest in front of the lobate salient of the thrust (Pl. 7) cannot be considered as evidence of the proportionate development of the salient in pre-Poison Canyon time, inasmuch as the shale doubtless flowed laterally as well as vertically. The probable effect of the shale on the structure is shown more diagrammatically in Figure 3. Surfaces of shearing in the shale at, and near, the contact between the Poison Canyon formation and the Pierre shale provide evidence in support of this interpretation of the structure. Also, the general crumpling and contortion of the shale in front of, as well as along the sides of, the prow-like lobe of the thrust leave no doubt as to the development of flowage and local thinning of the shale during deformation. The Pierre shale and other incompetent formations in the upper part of the section thus afford fundamental structural controls similar to those of the lower shales of the Pennsylvanian. The effect of these thicker incompetent formations is to cause underthrusting of more rigid plunger plates with respect to higher, and relatively rigid, buttress plates. This same action has been inferred with respect to the possibility of granitic plates underthrusting the higher Pennsylvanian and Permian conglomeratic beds along the western edge of the marginal belt (p. 957). The general effect produced is similar to that of a stratigraphic unconformity or disconformity, as was mentioned in the section on Pennsylvanian and Permian stratigraphy (p. 940).

DEPTH OF MARGINAL DEFORMATION

In attempting to visualize cross-sections of the belt of marginal deformation, not only must the direction of forces causing deformation be considered, but also the depth of deformation, which largely controls the forms of thrust plates and the relative amount of movement that may be inferred between each successive plate. It is, thus, necessary either to assume some depth of deformation or to assume some special form of thrusting. The use of the calculations made by Chamberlin in Colorado leads back to initial assumptions regarding the comparative depths of folding and thrusting, for thrusting forms an important element of the total shortening; if both folding and thrusting are involved in the total shortening, it must first be assumed that they are effective to approximately the same depth. In the marginal belt near Huerfano Park the dips of the thrusts range from vertical to horizontal, but on the whole are close to those of the adjacent folded formations, even where the faults themselves are, likewise, locally folded. Thus, it seems reasonable to assume that the faults of the marginal belt are really effective only to the depth of the folded strata to which they approximately conform.

Furthermore, it is essential to the usefulness of the results that the computation of depth of folding be restricted to the belt of folded and faulted rocks included in the marginal belt. The pre-Cambrian massif and its observable effect on deformation along the western border of the marginal belt should be discounted entirely, because of the probability that uplift of the massif is, in part, later than the folding and thrusting of the marginal belt. Evidence for this subsequent deformation will be presented later (p. 971).

It has been mentioned that the overlapping thrust plates of Huerfano Park, like the injective folds, contain no formations stratigraphically below the Pennsylvanian shales. The obvious conclusion is that the rocks involved in these two kinds of deformation have flowed over the basement upon which they lay. The lower series of shales presumably constituted the mobile layer within which the principal shearing and flowage took place. By calculating the depth of folding, one should obtain a check on the preceding conclusions, as the approximate thickness of the undisturbed formations near Huerfano Park is known.

Any calculation made must, thus, be based on initial postulates suggesting that, in the marginal deformation, only a shallow blanket of the crust is involved, and that folding and thrusting are effective to the same depth.

The section chosen for the following calculation is that near Bruff Creek in Huerfano Park, illustrated in Figure 2. Two interpretations of the

structure in the frontal overthrust plate are shown, for it is not certainly known whether the body of Permian beds at the east end overlie the Pennsylvanian in normal succession or belong to the overridden mass that lies beneath fault 'f3'. Little difference in measured shortening is obtained with either interpretation.

The reference or datum-plane used is within the Bridger and Huerfano beds, where it most closely fulfills the conditions outlined in Chamberlin's [43] calculation for northern Colorado—namely, that a minimum epeirogenic movement should have affected the orogenic disturbance, and that the ultimate base level to which the faulted and folded mountains are eroded should correspond as closely as possible to the level from which they arose.

The total shortening read from the structure sections ranged from 9.5 to 11 miles, and the calculated depth of folding appears to confirm the interpretation that the folding and faulting are shallow and involve only a plate not more than 2½ to 3 miles thick. This corresponds closely to the estimated total thickness of the formations in the vicinity of the section—hence, the underlying basement rocks are not necessarily involved in the deformation of the marginal belt.

DEFORMATION IN THE PRE-CAMBRIAN MASSIFS

GENERAL FEATURES

Pre-Cambrian crystalline massifs occupy a position adjoining the belt of marginal deformation, on the west, and range in width from the tapered end of a wedge at Orient to a maximum of 10 to 15 miles in the vicinity of the Sierra Blanca, a high mass of mountains in which lies Blanca Peak. At least two principal blocks of the crust are distinguishable: (1) the Sierra Blanca massif, an irregular wedge-like body, about 48 miles in exposed length, extending from the south slope of the Sierra Blanca northward to a point where it thins out just south of Orient; and (2) the Culebra massif, a block lying *en échelon* and overlapping the southern end of the Sierra Blanca massif, and forming the entire western part of the range, from Grayback Mountain into New Mexico. The Culebra massif is named from the older term given to the southern division of the Sangre de Cristo Mountains. The maximum exposed width of the Culebra massif in Colorado is about 8 miles; its length is about 48 miles, but it continues for some distance into New Mexico.

The forms of both massifs, as now exposed, appear to have been controlled by two important crustal disturbances, independent of each other both in time and in character, though possibly only representing different

[43] R. T. Chamberlin: *The building of the Colorado Rockies, pt. 2 (The Lyons-Grand River section)*, Jour. Geol., vol. 27 (1919) p. 235-236.

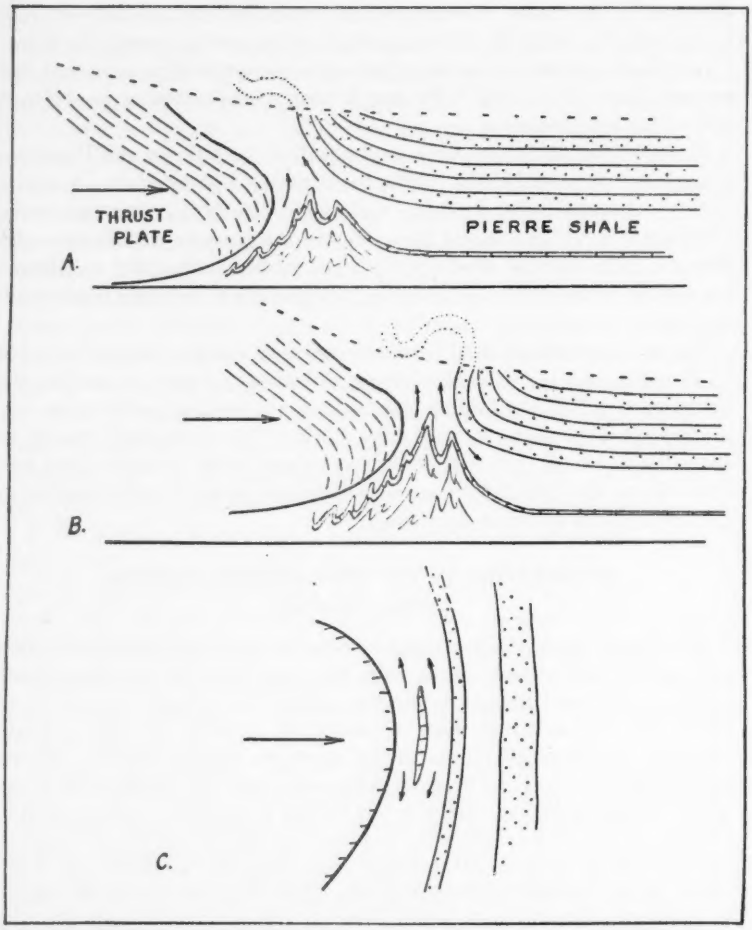


FIGURE 3.—*Diagrammatic illustration of plastic deformation in shale bodies near front of overthrust plates*

(A and B) Stages in the advance and rise of a lobate salient of a thrust plate. (C) Plan of lobate salient. Small arrows indicate probable direction of plastic flowage in Pierre shale.

phases of a major orogenic cycle. The earlier of these disturbances is that which has defined the eastern margins of the massifs and their contact with the belt of marginal deformation. Evidence for the belief that this movement represents the culmination in development of differential vertical forces set up in the shallower parts of the crust, and that this culmination

took place in late Bridger time, or later, will be presented. The later crustal movement is that which has defined the western margins of the massifs and is believed to be associated with the generation and extrusion of magma that resulted in the San Juan volcanic field and the compensating depression of the San Luis Valley. This disturbance probably occurred chiefly from late Oligocene to the end of Tertiary time. The main features of this deformation have already been described.[44]

CULEBRA MASSIF

The structure and nature of the rocks composing the southern part of the Culebra massif are practically unknown, except that the massif is supposed to be composed chiefly of pre-Cambrian rocks, partly concealed by Tertiary lavas along the west flank. In Plate 1, the boundaries of the geologic units are shown in dashed lines, to denote approximate and somewhat diagrammatic representation. It is known from scattered observations that the contact between this massif and the sedimentary formations on the east is, in general, steeply inclined or, possibly, a thrust contact with slight overturning toward the east. Within the body of the block, there are reported some infolded or infaulted bodies of sedimentary rocks. One early report, however, describing the lower Paleozoic formations on the west flank of the range northeast of San Luis has not been substantiated.[45] The western boundary of the massif is defined probably by a series of fault lines separating it from the downfaulted block of the San Luis Valley. These faults are reported to cut the Tertiary lavas.[46] The fault shown within the valley, and just west of the lava-covered San Pedro Mesa, is from an unpublished map by Whitman Cross and E. S. Larsen, and probably represents one of a series of faults along the western margin of the massif. The San Luis Hills, which lie west of this mesa, are described [47] as monadnocks, composed of volcanic materials and carved from tilted and uplifted fault blocks that have partly disrupted the late-Tertiary (Pliocene?) San Juan peneplain. In the vicinity of Fort Garland, mesas capped by gravel and lava are bounded, on the west, by escarpments that are approximately in line with the west front of the Sangre de Cristo Range. Regarding these physiographic features, Atwood and Mather state:

"Apparently a major fault plane with upthrow on the east runs approximately through Fort Garland in a direction a little west of north. Movement along this

[44] Whitman Cross and E. S. Larsen: *A brief review of the geology of the San Juan region of southwestern Colorado*, U. S. Geol. Surv., Bull. 843 (1935) p. 110-114.

[45] E. C. Van Diest, and P. H. Van Diest: *Notes on the geology of the western slope of the Sangre de Cristo Range in Costilla county*, Colo. Sci. Soc., Pr., vol. 5 (1898) p. 76-80.

B. S. Butler and C. W. Henderson: Personal communication to the effect that the supposed lower Paleozoic beds are really silicified lavas.

[46] E. S. Larsen: Personal communication.

[47] W. W. Atwood and K. F. Mather: *Physiography and Quaternary geology of the San Juan Mountains, Colorado*, U. S. Geol. Surv., Prof. Pap. 166 (1932) p. 99.

fault plane has in comparatively recent time lifted the Los Pinos gravel (Pliocene) above drainage level and caused its exposures in these mesas."[48]

The general physiographic and geologic evidence thus indicates that the Culebra and the southern end of the Sierra Blanca massifs are separated from the San Luis Valley block by faults that were active until late Tertiary or Quaternary times. As shown by the San Luis Hills, the downfaulted block within the valley, which elsewhere is widely covered with alluvium, is also broken into numerous small and tilted fault blocks by the forces that represent the last stage of deformation, and have caused a disruption of the formerly more extensive massifs.

In the central and northern parts of the mountains, particularly at the northern termination of the Culebra massif on Grayback Mountain, more evidence has been obtained regarding the earlier history of these massifs and their relation to the marginal belt of deformation. The northern part of Grayback Mountain consists chiefly of coarse-grained granite, which incloses small masses of hornblende gneiss. As shown in Plate 4, the mountain terminates northward in a blunt nose, about 1¾ miles wide. This nose of the mountain and the relations of the surrounding sedimentary formations to it indicate that the Grayback Mountain block is sharply upthrust from the basement, and is not the nose of a gradually plunging anticline. The surrounding sedimentary beds dip steeply away from the nose on all sides, and the sharpness in change of their strike at the corners corresponds to the sharp corners of the nose itself. Along the western side of Grayback Mountain the actual contact between the granite and the vertical sedimentary beds is exposed in several ravines. Wherever seen, the contact is an essentially vertical fault along which the lower Pennsylvanian beds are in contact with granite. On the eastern side of the mountain mass, near its north end, the actual dip of the granite wall is not exposed, owing to the absence of deep ravines and to a covering of Pleistocene gravels. The abutting sedimentary beds, however, have an easterly dip of 65 to 90 degrees. About 5 miles farther south, exposures in the Grayback mining district show that the east contact is here, likewise, nearly vertical and in places is a distinct fault.[49] The sedimentary beds flanking the short northwest face have a northeasterly strike, dip 40 to 60 degrees northwest, and are separated from the granite block by a zone of crumpling and faulting. Further evidence on the tectonic nature of the contacts is also afforded at the northwest end of the block, where a conglomeratic quartz-pebble facies of the basal Pennsylvanian beds rests with normal contact upon the top of the granite and gneiss. Four or five hundred feet above the bottom of the canyon of the Huerfano

[48] *Op. cit.*, p. 100

[49] H. B. Patton and others: *Geology of the Grayback Mining district, Costilla County, Colorado*, Colo. Geol. Surv., Bull. 2 (1910) p. 74, pl. 2.

River, these beds lie nearly horizontal or dip 15 to 20 degrees and are gently warped but turn down sharply along the northwest face. It is perhaps significant that no facies of sedimentation corresponding to these basal beds was found at the well-exposed contacts along the west side of Grayback Mountain. Where the automobile road crosses the east margin of the Culebra massif, west of Veta Pass, there are layers of conglomeratic beds somewhat more arkosic than those on Grayback Mountain. These appear to be in normal contact with the steep face of the granite, but considerable shearing and faulting is evident in the shales and sandstones that overlie the conglomeratic beds. The evidence presented by these exposures suggests that the northwest end of the Culebra massif, represented by the Grayback Mountain block, rose vertically through the overlying sedimentary blanket, carrying with it parts of the Carboniferous formations, which remained plastered on the top and sides. This interpretation is illustrated in Figure 4, a diagrammatic cross-section of the Culebra and the Sierra Blanca massifs.

Two fundamental concepts of the origin of the Grayback Mountain block are possible: first, that it was an autochthonous body of the basement immediately underlying the zone of overthrusting and was forced upward through an overlying blanket of deformed beds in the marginal zone; second, that the block is a body torn from the basement and carried eastward by the overthrusting that caused deformation in the marginal zone. The rectilinear fault faces are difficult to associate with the idea of overthrust fault blocks, and, in this respect, the body differs from the smaller overthrust bodies of pre-Cambrian granite in the Kerber Creek region, which are lobate in outline. The possibility that the rectilinear fault faces are the result of steep faulting across an older, and formerly more extensive, overthrust sheet is not believed to be in harmony with the outline of the Culebra massif as a whole, of which the mountain is a part, unless it be assumed that the Culebra massif is largely a thick overthrust mass. This interpretation is believed to be too extreme in its implications. There is, furthermore, evidence presented by the internal structure of the Sierra Blanca massif to show that differential vertical deformation affecting Carboniferous rocks has occurred within these larger pre-Cambrian bodies. This evidence conforms more closely with the upthrust origin of the massifs as a whole, though they evidently did not rise as a monolith like the comparatively small body of Grayback Mountain.

The Grayback Mountain block is separated from the Sierra Blanca massif, at the point where they overlap *en échelon*, by a trough about 2 miles wide, within which lie steeply dipping and strongly deformed Carboniferous beds. This trough narrows southward toward the valley of Ute Creek, which drains the southeast flank of the Sierra Blanca. Intensity

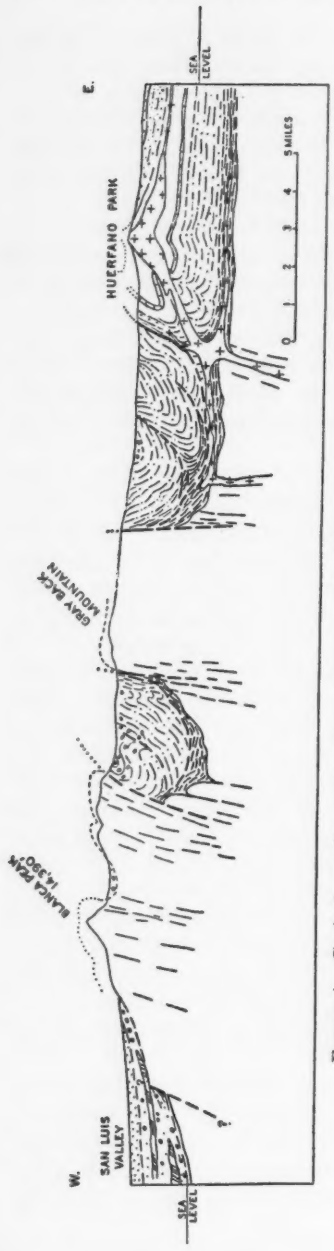


FIGURE 4.—*Geologic section across Sierra Blanca and Culebra massifs to the edge of Huerfano Park*

Section (along line A-A' of Plate 4) illustrates diagrammatically the probable late origin of the upthrust massifs, following a phase of thrusting and folding of the sedimentary blanket. Deformation in the sedimentary blanket must be largely accounted for by thrust plates of granite, such as those shown in Figure 1-B.

of deformation appears to be increasing in this direction, as some beds of limestone are converted to marble. This body of sedimentary beds was not studied in the Ute Creek area during the current work; hence, its southern continuation is known only from reports of the occurrence of sedimentary rocks there. The contact between the Sierra Blanca block and the Carboniferous beds crosses the saddle ridge between Blanca Peak and Grayback Mountain, but, unfortunately, that point was not reached during a reconnaissance trip into the area. Because the contact is elsewhere effectively concealed by high terrace gravels, Tertiary lavas, and debris from the mountain slopes, its dip is not accurately known. The beds along the west side of the trough dip from 45 to 60 degrees east but locally are much sheared and contorted. It appears, therefore, that the contact is tectonic, like that in the section east of Crestone, and that the Blanca Peak block may have overridden the trough at its west margin along a steep zone of shearing. The strong compression within this trough separating the two major massifs, and the apparent increase in metamorphism as the trough narrows southward, correspond,

in some respects, to minor structures, to be described, within the Sierra Blanca massif itself.

SIERRA BLANCA MASSIF

As mentioned in the section on the pre-Cambrian rocks, the Sierra Blanca massif is composed chiefly of granite, diorite, hornblende gneiss, and injected schists and gneisses. The strike of gneissic and schistose rocks in the Sierra Blanca Mountain group is from northeasterly to east-west. In general, therefore, in its southern part, the strikes of the rocks composing the massif do not correspond with the outlines of that body. Both this fact and the high degree of metamorphism exhibited by these rocks show that they belong, without doubt, to the pre-Cambrian complex. Actually, several narrow slices of highly compressed Carboniferous sandstones and shales have been found confined between essentially vertical walls of the pre-Cambrian complex. One such slice was found near the sharp bend in the eastern boundary of the massif, about 15 miles north of Blanca Peak; it is several thousand feet long but only a few hundred feet wide, and is quite inconspicuous. B. S. Butler has reported to the writers the occurrence of another body of this kind on the west slope of the range, southeast of Crestone. The strikes of the slices are essentially parallel to the general trend of the massif.

The writers were unable to determine whether the slices in the Sierra Blanca massif represented the basal Pennsylvanian beds caught between vertically imbricated blocks of the basement, or some chance section of Carboniferous beds entrapped by differential plastic deformation in the basement rocks on either side. As plastic deformation has occurred in the Carboniferous beds and locally converted them to schists, it is probable that such deformation has also taken place in deeper parts of the basement rocks beneath the mountains—parts that have since been raised above the zone of plastic flow. The entrapped slices give visible evidence of greatly increased induration, for their sandy beds now have the appearance and hardness of quartzite. The solution of these related problems must obviously await more critical study.

With these examples in mind, it may be conceived that the trough lying between the Sierra Blanca and Grayback Mountain is a similar structure, entrapped by the upward expansion of the massifs on either side, the deformation within the trough being intensified by some lateral expansion of the much larger Sierra Blanca block. As was indicated in the introduction to this section, it is believed that the Sierra Blanca massif was originally much wider throughout its length, its present western limits having been produced by later downfaulting of the San Luis Valley block. Further strong evidence in support of the hypothesis that the Sierra Blanca massif has been disrupted is found at its northern end. Between the Huerfano Park region and Crestone the eastern contact of the mas-

sif has not been traced, but from Crestone north to its end the contact has been examined. Opposite Crestone the east contact dips about 75 degrees west, and the Carboniferous beds east of it are considerably deformed (Fig. 1, A). Nearer the exposed north end of the massif the lower Paleozoic beds along the contact are overturned with dips of 40 to 60 degrees westerly. Close to the north end, there are overthrusts of even lower angle, along which the overturned lower Paleozoic beds have been shoved eastward over the Pennsylvanian beds. The faults and overthrust beds, however, turn northwestward, striking obliquely to the mountain front, and are truncated by the linear boundary between the mountains and the valley. This thrust zone reappears in line with its strike on the west side of the valley, in the Kerber Creek region, the faults of which have been previously described by Burbank.[50] Here, the main contacts between the granite and the overridden Carboniferous dip 35 to 50 degrees southwest to south. Thus, there appears to be a gradual decrease in the dips of the latest thrust planes bounding the Sierra Blanca massif toward its northern end.

Another feature of the structure near Kerber Creek is the occurrence of klippen or outliers of a thrust lying from half a mile to a mile and a half north of the steep thrust planes bounding the massif (Pl. 1). The sole of the thrust upon which these granite klippen lie is characterized by a breccia zone, 20 to 30 feet thick, within which fragments of the underlying Carboniferous beds have been partly comminuted between the granite blocks. The attitude of this outlying thrust surface is irregular but is generally horizontal. The nature of the thrust surface, as already described, contrasts greatly with that of the steeper thrust surfaces bounding the granite massifs to the south, as the steeper thrusts have inconspicuous striated faces, with evidence of shearing in both the granite and the sedimentary beds near the contact. This shearing in the Kerber Creek area, however, is less intense, and there is less plastic deformation, as compared with the contacts of the Sierra Blanca massif near Crestone. One may perhaps infer that the boundary of the massif exposed near Kerber Creek represents a shallower phase of deformation. This inference would be further supported by such facts as the presence of only moderately tilted lower Paleozoic beds resting upon the granite bodies south of the thrust zone, as contrasted with the steeply inclined and sheared slices of Carboniferous beds in the more southern parts of the massif. Carrying the implications of these inferences further, one might deduce that the thrust planes bounding the Sierra Blanca massif become steeper at depth. Actual observation at the contact near Crestone gives some suggestion of this steepening, although the evidence is not conclusive.

[50] W. S. Burbank and C. W. Henderson: *Geology and ore deposits of the Bonanza mining district, Colorado*, U. S. Geol. Surv., Prof. Pap. 169 (1932) p. 38-41, pl. 3.

The klippen in the Kerber Creek region have been interpreted as representing parts of a low-angle thrust, the development of which preceded that of the higher-angle thrusts that bound the granite massif.[51] The presence of such low-angle thrusts or overthrust sheets in the northern part of the massif suggests that a search be made for some evidence of a corresponding phase of thrusting in the more southern portion. As the southern part of the massif appears to have been uplifted relatively more than the northern end during later stages of deformation, it is possible that the sole of the overthrusting, corresponding to the klippen at Kerber Creek, now lies at greater depth in the southern end of the massif and, hence, corresponds to the already described marginal phases of deformation east of the massifs. Thus, one may picture the early stages of marginal deformation as having been produced by sheets of granite, thrust out from the edges of the Sierra Blanca massif and overriding the marginal zone, crumpling and imbricating the sedimentary blanket in front of them. This interpretation has been shown diagrammatically in Figures 1, B, and 3. Such an interpretation would be difficult to confirm other than to add to, and complete, the kind of evidence already presented, and other explanations may fit these facts equally well. The downfaulting and tilting of the San Juan Mountain block and its concealment by the San Luis Valley fill prevent a complete study of the inner structures of the Sierra Blanca and the Culebra massifs. However, if thrust sheets of this kind overrode the tops of the massifs, possibly some deformed remnants of them may yet be preserved in the higher unmapped summits, such as Blanca Peak.

CORRELATION OF STAGES WITHIN THE MASSIFS

Deformation in the pre-Cambrian massifs thus appears to have two recognizable stages: an early stage, represented by the klippen at Kerber Creek, which may be remnants of thrust plates of granite that overrode the border zone of the Sierra Blanca massif; and a later stage of upthrusting, which defines the present eastern edge of the massif and is characterized by comparatively high-angle thrusts that probably steepen in depth. The early stage is tentatively correlated with deformation in the marginal zone of the Sangre de Cristo Range, which occurred in pre-Huerfano time. The great uplift of the massifs, with its accompaniment of plastic deformation along their borders and within them, seems to be a later event, as shown by the structure of Grayback Mountain and by the relation of the steeper thrust planes to the klippen at Kerber Creek. This uplift calls for a maximum vertical movement of about 3 miles in Blanca Peak and about 2.5 miles in Grayback Mountain, as shown by the hypo-

[51] W. S. Burbank and C. W. Henderson: *op. cit.*, p. 41.

thetical section of Figure 4, though the amount of this movement assignable to post-Huerfano time might be considerably reduced, if it could be shown that upwarping of these blocks was initiated during the earlier phases of marginal thrusting.

The distribution of the uplift between pre- and post-Huerfano movements can be approximated, on the assumption that the uplift of the Greenhorn Mountains, comprising the southern termination of the Wet Mountain massif, occurred at the same time as that of the Sierra Blanca and the Culebra massifs. With regard to the Greenhorn Mountains, there is a fair amount of evidence showing that this movement attained great proportions in late or post-Bridger time, although gradual uplift took place between the early overthrusting and the post-Bridger movement. This statement paraphrases the conclusions of R. C. Hills,[52] who at one time attempted to show that the effects of pre-Bridger and post-Laramie movement in Colorado "have been often overrated, and results ascribed to it should have been credited to succeeding orogenic disturbances." Hills was obliged later to modify these conclusions to some extent, for in his earlier work, he had failed to recognize the unconformities existing between the Poison Canyon and the Cuchara and between the Cuchara and the Huerfano formations. As the Poison Canyon formation shows marked unconformity in its relation to the underlying Cretaceous rocks, even along the flank of the Greenhorn Mountains, it must be conceded that this uplift was initiated during the earlier stages of overthrusting that caused the complex deformation of the marginal belt. The upturning of the later Huerfano beds along the west flank of the Greenhorn Mountains forms a monocline, terminated by sharp upturning of the Cretaceous and older beds close to the steep bounding fault of the range.

Though the Huerfano Basin was perhaps initiated in early Eocene time, the character of the debris deposited in this lowland in post-Poison Canyon time shows that the source beds were chiefly the Carboniferous formations of the low mid-Eocene Sangre de Cristo Mountains. The evidence for a comparatively moderate relief and warm temperate climate, shown by the lithology and fossil remains of Bridger time, has been mentioned in preceding paragraphs (p. 947). The beds of Bridger age, which comprise the upper portion of the Huerfano formation, occupy part of a zone ranging from 2000 to 2500 feet above the base of the formation. Very little coarse granitic debris is to be found in the Bridger sediments of Huerfano Park; hence, the Huerfano beds either actually covered the neighboring granitic core of the Greenhorn Mountains, on the one hand, or, on the other, the Greenhorn axis at this time was still covered with a thin layer of Carboni-

[52] R. C. Hills: *Orographic and structural features of Rocky Mountain geology*, Colo. Sci. Soc., Pr., vol. 3 (1890) p. 362-458.

ferous strata, upon which the Huerfano beds may have overlapped. Furthermore, the lower Huerfano beds are now eroded and flexed along the Greenhorn uplift, though there is no evidence of unconformity within the formation. Doubtless, as Hills [58] has suggested, a considerable part of the uplift of the Greenhorn axis is of late- or post-Bridger age. The top of the pre-Cambrian core of the range in mid-Eocene time, thus, probably stood either lower than, or close to, the level of the top of the Huerfano formation. At the present time, however, the pre-Cambrian core of the range stands at an elevation of about 11,500 feet above sea level, as compared to an elevation of 8700 feet for the top of the Huerfano formation; hence, a differential uplift of not less than 2800 feet seems to have occurred since Bridger time. If, however, Carboniferous and lower Paleozoic beds had not been entirely eroded from the basement in Bridger time, a somewhat greater uplift would be called for. A greater uplift would also be necessary under the assumption that the relatively mature surface now forming the tops of the mountains and partly covered with Tertiary lavas was produced by considerable post-Bridger erosion of the pre-Cambrian core. Such post-Bridger erosion seems improbable. Rather, it would seem that the removal of the sedimentary cap occurred before or during the deposition of the Cuchara and earliest Huerfano beds, for the pre-Cambrian basement was protected during middle and later Eocene by a partial covering of the Huerfano beds and during middle and late Tertiary by a covering of lavas. However, a short interval of erosion of pre-Cambrian rocks appears to have intervened between early Tertiary time and the mid-Tertiary period of volcanism. Thus, in the Wet Mountain Valley just north of Huerfano Park, there is a thin formation of mixed clayey and granitic material, resting directly upon pre-Cambrian granite and underlying the volcanic beds. In sharp contrast, conglomerates in the volcanic formations are remarkably free of granitic material.

In conclusion, it is probable that not less than 2800 feet of uplift of the Greenhorn Mountain axis relative to the Huerfano Basin has occurred since early Bridger time. If the stripping of the Paleozoic beds from the mountain axis chiefly antedated Huerfano time and the surface formed represents a re-exhumed peneplain, upon which the Cuchara and early Huerfano beds rested, a maximum relative uplift, not to exceed about 4500 to 5000 feet, could have occurred. This figure is slight as compared with the maximum relative uplifts of 2 to 3 miles, estimated for the Culebra and the Sierra Blanca massifs. As in the case of the Greenhorn uplift, the Culebra and the Sierra Blanca massifs were probably partially uplifted as crustal folds prior to their final upthrust in post-Huerfano time.

[58] R. C. Hills: *Description of the Walsenburg quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 68 (1901) p. 2-3.

If the late uplift of the Greenhorn Mountains was synchronous with that of the Sierra Blanca and the Culebra massifs, there appears to have been a decrease from west to east in the intensity of orogenic forces causing this deformation. Though less study has been given to overthrusts along the east margin of the Wet Mountain uplift, it is known that there is not a wide marginal zone of overthrusting and that the angle of thrusting is generally steeper than that in the marginal zone of the Sierra Blanca. These differences in the nature of deformation in the two ranges may, perhaps, be attributed to changes in the physical state of the shallow crust, rather than to local changes in the application of the forces involved. Thus, the higher-angle thrusting along the eastern margin of the Sierra Blanca and of the Wet Mountains may represent a comparatively rigid state of the shallow crust, as compared with the more plastic extrusion of low-angle thrust plates, which seem to have been confined chiefly or entirely to the western massifs.

The stage resulting in the final disruption of the Sierra Blanca and the Culebra massifs by late Tertiary time, however, is believed to represent a different cycle of deformation, as it is characterized by tensional faulting and foundering of the greater part of the massifs beneath effusions of lava. The transition between the early upthrusts and this subsidence of the crust must have occurred between late Eocene and late Oligocene times.

IGNEOUS ACTIVITY IN RELATION TO STRUCTURAL HISTORY

Igneous rocks have been mentioned only incidentally in the preceding pages, but a brief summary will serve to give a better general idea of the role of igneous activity in relation to the structural history. All the known Tertiary intrusive bodies are in the sedimentary rocks of the mountains. In the northern range the igneous rocks are found chiefly as medially situated bodies of small size near, or within, the central anticline of Pennsylvanian sedimentary rocks. These bodies are composed, in part, of quartz diorite or granodiorite and, in part, of porphyritic equivalents of these rocks. With them are associated small bodies of monzonitic, granitic, and pegmatitic differentiates, and dikes of felsite or aplite. The intrusive body east of Orient (Pl. 1) conforms in its elongate shape and steep contacts with the strike and dip of the Pennsylvanian beds of the central anticline. Exposures of similar bodies at fairly regular intervals from the vicinity of Orient south at least to Crestone suggest that at moderately greater depths the magma may have been fairly continuous lengthwise of the range. The structural relation of the shallower and inferred deeper bodies is shown diagrammatically in Figure 1, A.

South of Huerfano Park, most of the exposed igneous bodies lie along the far east flank of the marginal belt (Pl. 1; Fig. 4) in Cretaceous and Ter-

tiary formations, though some are close to the eastern edge of the Culebra massif in the Carboniferous beds. The intrusive bodies in the older formations are best illustrated by the examples of the Grayback mining district.[54] The intrusive rocks are described as monzonite porphyry, quartz porphyry, and felsite, and comprise, for the most part, concordant injections.

The intrusive bodies of the eastern margin of the range extend from Huerfano Park south to the latitude of the Purgatoire River. They exhibit a wide range in composition from granite and syenite to basalt, and are especially noteworthy for the abundance of lamprophyric types. Some of the dikes and the intrusions of Spanish Peaks penetrate the Huerfano formation and, hence, may correspond in age to the later stages of mountain uplift.[55]

The extent to which the igneous magma was involved in the earlier stages of deformation cannot be easily determined, as the ages of intrusive bodies in the older formations are unknown. The presence of igneous rock fragments in the Cuchara formation suggests the existence of minor volcanism at, or prior to, mid-Eocene time. On the other hand, the general absence of much volcanic debris in the late Cretaceous and early Tertiary formations of the Sangre de Cristo region contrasts with other regions of Colorado. This suggests perhaps either an effective blanketing of magmas by the abnormally thick and strongly deformed sedimentary rocks, or the absence of much magma during earlier stages of deformation. Especially in view of the known long range of igneous activity throughout Colorado, beginning with the earliest stages of the Laramide orogeny, it may be concluded that blanketing of the magmas resulted in restricted volcanism even though magma may have been involved in earlier and deeper phases of deformation. It is noteworthy at least that the latest known intrusive rocks of the Spanish Peaks area show a much more strongly contrasted series of differentiates than those exhibited by the central intrusive bodies of the mountain ranges, which belong perhaps to an older group.

As shown in Plates 1 and 4, lava beds of mid- to late-Tertiary age flank both sides of the Sangre de Cristo Mountains. The sources of the lavas of the Huerfano River area are not known, but they are presumably only small remnants of originally widespread accumulations that partially buried the mountains during final stages of tensional faulting.

CONCLUSION

The writers believe it is inadvisable to make any broad or final conclusions concerning the orogenic history of the Sangre de Cristo Mountains.

[54] H. B. Patton and others: *Geology of the Grayback mining district, Costilla County, Colorado*, Colo. Geol. Surv., Bull. 2 (1910) p. 35-49.

[55] R. C. Hills: *Description of the Spanish Peaks quadrangle*, U. S. Geol. Surv., Geol. Atlas U. S., fol. no. 71 (1901) p. 3-4.

The work done by Hills, Melton, Johnson, and the writers has established some of the general features as well as the local complexity of the geologic structure, and brings out the necessity of extreme care in coordinating further studies of the geologic structure and the stratigraphy. Past studies of the range have, like the present one, been handicapped by lack of key horizons in the Pennsylvanian and Permian formations, which constitute much of the complexly faulted and folded parts of the mountains.

The principal purpose of this paper is to emphasize the need of more systematic investigations in this region, and to point out that any one of the general hypotheses of mountain-building that have been applied to the southern Rocky Mountain province fails to take into account possible changing phases of deformation during the complete span of the "Laramide revolution".

Places particularly favorable for further study are indicated in the text, but it would seem that detailed mapping of the Sierra Blanca and Huerfano Park regions will probably yield more conclusive evidence as to the age and nature of various phases of orogeny. The southward continuation of the Culebra massif in New Mexico will also be likely to afford better exposures of the inner structures of the massifs, because of their partial concealment in Colorado beneath the lavas and Tertiary sediments of the San Luis Valley. Despite Hills' earlier work, which established the existence of the later Eocene formations bordering the belt of deformation, neglect of this region has resulted in too much emphasis upon later stages of uplift in the southern Colorado province.

U. S. Geological Survey, Washington, D. C.
Manuscript received by the Secretary of the Society, July 1, 1936.
Read before the Geological Society, December 31, 1931.
Published with the Permission of the Director of the U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 48, PP. 977-1000, 11 FIGS. JULY 1, 1937

STRUCTURE OF THE HONEYBROOK UPLIFT, PENNSYLVANIA

BY GEORGE W. STOSE

CONTENTS

	Page
Introductory statement	977
General structure of the region	978
Welsh Mountain anticline	981
Barren Hills anticline	991
Mine Ridge anticline	995
Later cross faults	999
Age of the structures	999

ILLUSTRATIONS

Figure		Page
1.	Generalized map of southeastern Pennsylvania	979
2.	Geological map of the Honeybrook uplift	980
3.	Geological map of the eastern part of the Welsh Mountain anticline	983
4.	Geological map of the western end of the Welsh Mountain anticline	986
5.	Map of the vicinity of Brandywine Manor	987
6.	Structure sections showing hypothetical stages of uplift of the Welsh Mountain anticline	989
7.	Maps showing hypothetical stages of uplift of the Welsh Mountain anticline	990
8.	Geological map and section of the south edge of the Barren Hills anticline at Devault	993
9.	Generalized geological map of the western part of the Mine Ridge anticline	994
10.	Maps of the Mill Lane anticline	996
11.	Theoretical interpretations of the Mill Lane structure	997

INTRODUCTORY STATEMENT

The Honeybrook uplift lies on the east side of the large Paleozoic limestone lowland around Lancaster in the Piedmont of southeastern Pennsylvania. This lowland is an embayment of the limestones of the Appalachian Valley into the Piedmont area around the plunging end of folds of Cambrian quartzites of South Mountain, which is the northern extension of the Blue Ridge (Fig. 1). The Honeybrook uplift brings Cambrian quartzites and pre-Cambrian basement rocks to the surface on the east side of the limestone lowland. Chester Valley, a long narrow synclinal limestone valley of east-northeast trend, lies south of the Honeybrook uplift, which is overlapped on the northeast side by gently north-dipping Triassic redbeds.

This paper is the result of detailed studies for a report on the Honeybrook and Phoenixville quadrangles to be published by the United States Geological Survey. The pre-Cambrian rocks, which were mapped by Florence Bascom and will be fully described by her in the forthcoming report, are here treated as a unit. They include metamorphosed sedimentary gneisses and intrusive igneous gneisses, similar to those in the Reading Hills and the Highlands of New Jersey.

TABLE 1.—*Paleozoic sedimentary rocks*

Age	Name	Character	Thickness (feet)
Lower Ordovician	Conestoga limestone	Thin-bedded argillaceous limestones and purer crystalline limestone of Beekmantown age, in Chester Valley.	500–800
	—Unconformity—		
Middle Cambrian	Elbrook limestone	Fine-grained earthy limestone and marble.	800±
Lower Cambrian	Ledger dolomite	Thick-bedded pure granular dolomite.	1000±
	Kinzers formation	Thin dark shale and nodular banded argillaceous limestone.	130–30 (C. V.)*
	Vintage dolomite	Thick-bedded dark-blue knotty dolomite.	600–200 (C. V.)
	Antietam quartzite	Thin-bedded ferruginous quartzite.	450–100 (C. V.)
	Harpers phyllite	Light-gray to dark quartzose finely micaceous phyllite.	800–500 (C. V.)
	Chickies quartzite Hellam conglomerate member at base	Scolithus-bearing vitreous quartzite, conglomeratic at base.	1000–500 (C. V.)

* The formations below the Ledger dolomite are all thinner in Chester Valley, as shown by the figures accompanied by C. V.

The sedimentary formations involved in the structures to be described are Cambrian quartzites, and Cambrian and Ordovician limestones. They are listed and briefly described in Table 1. The Triassic sedimentary rocks and thick intrusive diabase that border the area on the north are mentioned only where they are locally included in the structures described. Diabase dikes also cut the Paleozoic rocks.

GENERAL STRUCTURE OF THE REGION

The Honeybrook uplift includes three distinct anticlines—Welsh Mountain, Barren Hills, and Mine Ridge anticlines (Fig. 2). They rise out of a limestone lowland, which surrounds them on the north, west, and south, and are overlapped on the north and northeast by Triassic rocks.

The limestone region to the west of the uplift is a rolling cultivated lowland about 450 feet in general altitude, into which the streams have cut open valleys 100 to 150 feet deep. Chester Valley, to the south, is about 350 to 550 feet in general altitude, and the streams have cut valleys 50

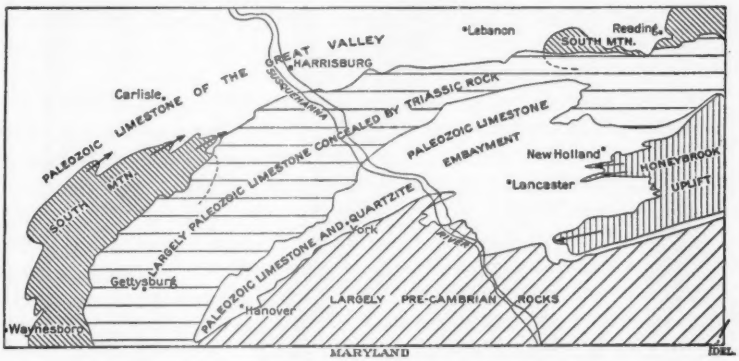


FIGURE 1.—*Generalized map of southeastern Pennsylvania*

Showing relation of Honeybrook uplift to the Paleozoic limestone embayment in the Piedmont province. Arrows show pitch of folds of the Cambrian quartzites in the uplifts.

to 100 feet deep into this lowland. A rather broad, level surface on the pre-Cambrian rocks around Honeybrook is about 750 feet in altitude, above which rise mountains composed of Cambrian quartzite. Welsh Mountain is a ridge with several tops more than 1,000 feet in altitude, the highest reaching 1,080 feet. The Barren Hills rise to a maximum of 960 feet, and Mine Ridge to 920 feet.

The area here described lies largely in the Honeybrook and Phoenixville quadrangles. The Welsh Mountain anticline extends westward into the New Holland quadrangle, the geology of which has been published.[1] In that report the structure was not satisfactorily described nor were the structural problems fully solved. The Mine Ridge anticline extends southwestward into the Coatesville quadrangle[2] and west of the Coatesville into the Quarryville quadrangle,[3] the geology of which has also been published. The generalized geology of the area is shown in Figure 2.

For those who have not worked in this part of the Appalachian and, therefore, are not familiar with conditions in the field, it will be helpful

[1] Anna I. Jonas and George W. Stose: *Geology and mineral resources of the New Holland quadrangle, Pennsylvania*, Pa. Geol. Surv., 4th ser., Top. and Geol. Atlas, no. 178 (1926) 40 pages.

[2] F. Bascom and G. W. Stose: *Geology of the Coatesville and West Chester quadrangles*, U. S. Geol. Surv., Geologic Atlas U. S., fol. 223 (1932) 15 pages.

[3] E. B. Knopf and A. I. Jonas: *Geology of the McCalls Ferry and Quarryville quadrangles, Pennsylvania*, U. S. Geol. Surv., Bull. 799 (1929) 156 pages.

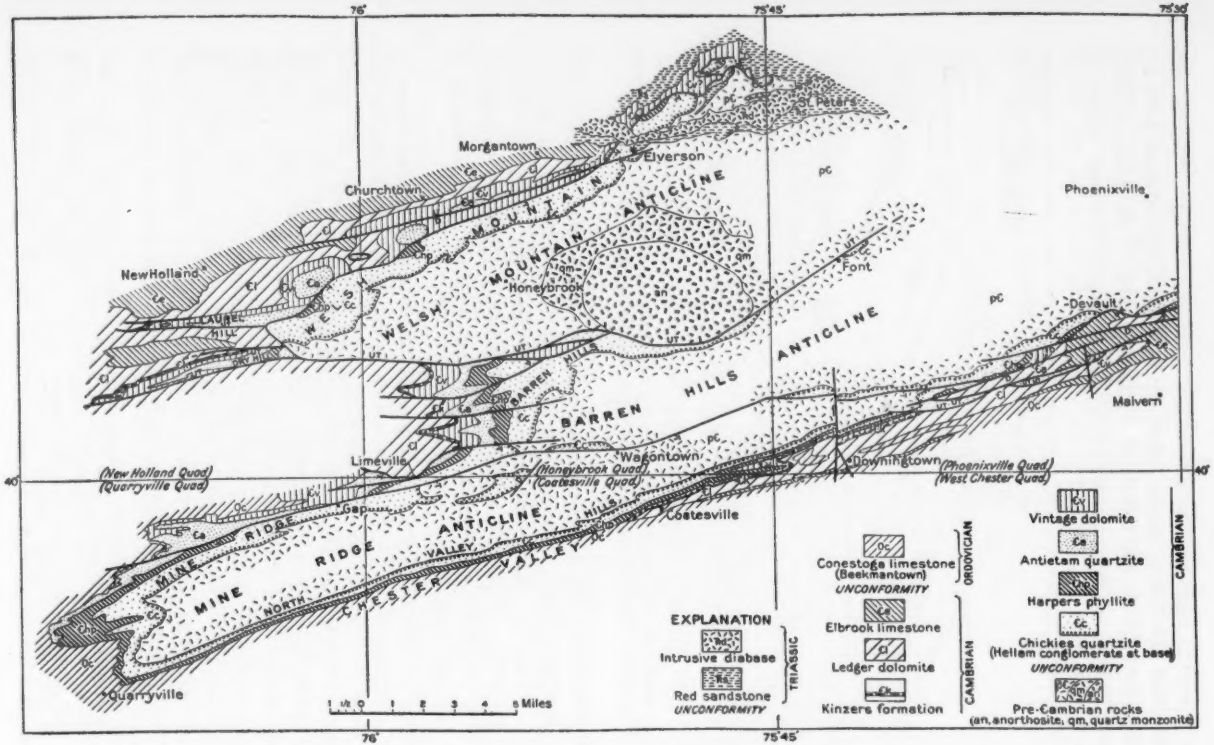


FIGURE 2.—*Geological map of the Honeybrook uplift*

Showing three major anticlines. (T) Overthrust side of low-angle thrust fault, (UT) upthrust side of high-angle fault, (D) downthrow side of normal tension fault, (Arrow) direction of horizontal movement of shear fault.

to state that faults are seldom exposed because of deep soil and weathered outcrops, and the exact location of the break, the dip of the fault plane, and the character of the rocks adjacent to the fault are generally not known. The presence of a fault is recognized by the fact that there is a break in the normal sequence of formations that can be explained best by the assumption of a fault. Deep weathering of the rocks, the presence of iron-ore deposits, springs, vein quartz, or ferruginous chert help to locate the fault. In places, a fault breccia, cemented by iron oxide, suggests an open break or normal tension fault. In other places, a shear zone or a mylonite indicates faulting under compression. Upthrow of the rocks on the south side of a fault suggests an overthrust, for most of the folds in this part of the Appalachians are overturned to the northwest, and the overthrusts have moved in that direction. If the upthrow is on the north side, the fault is likely to be a normal tension fault and not an overthrust. The attitude of the rocks on both sides of the break, the relation of the fault to the topography, and the rock structure, if observable, are an aid in deciding the nature of the fault movement. It is evident that in solving the structural problems, especially the character of the faults, under such conditions, conclusions must be reached with insufficient evidence. Some of the structural interpretations and conclusions in this paper, therefore, may be stated more positively than may seem warranted.

WELSH MOUNTAIN ANTICLINE

The Welsh Mountain anticline exposes a core of pre-Cambrian rocks, with Cambrian quartzites on its north limb. These quartzites form Welsh Mountain, a high ridge, 12 miles long, and other detached hills to the northeast along the strike of the rocks. The pre-Cambrian core has a width of 7 miles east of Honeybrook.

The Cambrian quartzites on the north limb of the anticline dip north, except where they are locally overturned, as on Welsh Mountain, above Narvon, where the quartzite dips 80° to 55° south. The quartzite near the mountain front is sheared in places to a sericitic quartz mylonite, the schistosity dipping 35° to 40° south. The schistose quartzite has been leached in places by circulating waters and reduced to white sericitic clay, extensively mined around Narvon. This schistose zone indicates a thrust fault, which is difficult to trace along the north foot of the mountain because of the thick cover of quartzite roll and wash from the mountain slope. The absence of the Antietam quartzite and Harpers phyllite along most of this contact and the presence of many old iron-ore banks and pits are regarded as further evidence of the fault. The fault is placed at the foot of the steep slope where the ore pits are. The Chickies quartzite is thrust over Vintage dolomite in the lowland along the whole length of

Welsh Mountain as far east as Elverson (Figs. 2 and 4), and the fault is named the Elverson overthrust.

The Elverson overthrust is, in general, a strike fault and follows the north foot of the Chickies quartzite ridge except where it curves around the ends of minor anticlines and cuts into the overlying Harpers phyllite in minor synclines, as shown southeast of Beartown. Its trace on the surface is especially sinuous in the western part of the uplift, where it follows around the plunging end of larger folds. Several anticlines in the limestone in front of the overthrust bring Antietam quartzite to the surface, and some of these anticlines are overridden by the fault. The Chickies quartzite is thrust over an anticline of Antietam quartzite, which forms the prominent outlying oval hill northeast of Beartown, and two smaller anticlines of Antietam near Elverson are similarly overriden by the quartzite (Figs. 2, 3, and 4). The prominent oval hill south of Cedar Lane, near the west end of Welsh Mountain, is another anticline of Antietam quartzite in front of the overthrust. Laurel Hill, which extends westward in the limestone lowland 3 miles beyond the end of Welsh Mountain, is an elongated anticline of Antietam, which strikes under the overthrust mass of Chickies quartzite and Harpers phyllite (Fig. 4). Between Beartown and Cedar Lane the Chickies quartzite is overthrust on Ledger dolomite, and a low hill of hard crushed quartzite of Chickies type, in the limestone lowland southwest of Beartown, is interpreted by the writer as a remnant (klippe) of the overthrust quartzite.

Southwest of Elverson the overthrust is cut and offset by a nearly east-west normal tension fault of later date. This fault extends westward for several miles, and the upthrow on its north side brings to the surface two anticlinal areas of Antietam quartzite, which make low ridges in the limestone lowland southeast and southwest of Morgantown (Fig. 2). A minor fold in the overthrust mass southwest of Elverson raises the basal Hellam conglomerate member of the Chickies quartzite to the sole of the overthrust, and the conglomerate here overrides one of the small anticlines of Antietam quartzite in the limestone (Fig. 3). North of the Elverson, another break in the overthrust is caused by a thick Triassic diabase dike, which cuts across the structures in the Paleozoic rocks and passes into the overlying Triassic sedimentary beds. The rocks on the northeast side are uplifted by the intrusion, and the gently dipping Triassic redbeds are offset northward about 1 mile, but the structures in the Paleozoic rocks are not markedly affected.

Northeast of Elverson the Chickies quartzite on the north limb of the anticline is not so continuous. It forms Thomas Hill and several other hills to the northeast (Fig. 3). An anticline of Antietam quartzite north

of the overthrust is exposed in the lowland west of Thomas Hill, 1 mile north of Elverson, and Vintage dolomite is believed to be present beneath

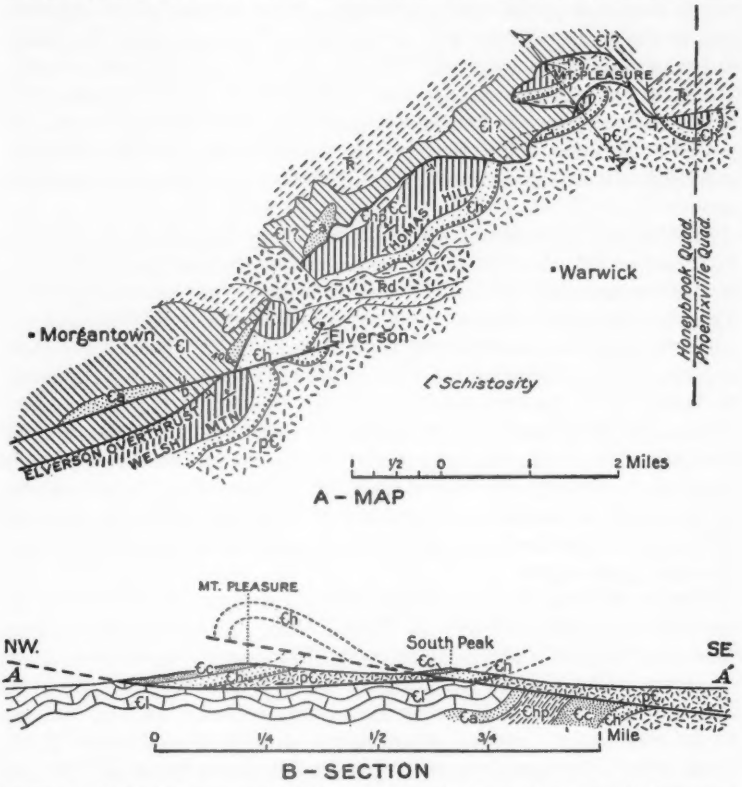


FIGURE 3.—*Geological map and section of the eastern part of the Welsh Mountain anticline*

(A) Geological map showing the Elverson overthrust, which is broken by a diagonal normal fault and by a thick Triassic diabase dike. At its east end the overthrust splits into two slices. Minor folds in the overthrust mass bring up the Hellam conglomerate and pre-Cambrian rocks to the sole of the overthrust. The location of the fault in the lowland north and northeast of Thomas Hill is conjectural. (B) Cross-section through Mt. Pleasure, along line A-A', showing split of the Elverson overthrust into two slices. T_R, Triassic sedimentary rocks; Trd, Triassic diabase dike; €l, Cambrian limestones; €a, Antietam quartzite; €hp, Harpers phyllite; €c, Chickies quartzite; €h, Hellam conglomerate member; p€, pre-Cambrian rocks; T, overthrust side of low-angle fault; D, downthrow side of normal tension fault; U, upthrow side of normal fault.

the alluvium in the lowland surrounding the Antietam, although it is nowhere exposed. The trace of the fault here is, therefore, conjectural.

Another minor fold in the overthrust mass northeast of Thomas Hill brings the Hellam conglomerate, and probably the underlying pre-Cam-

brian rocks, to the sole of the overthrust, producing a marked break in the quartzite ridge. In the lowland northeast of Thomas Hill the structure is concealed by alluvium and is further complicated by an apparent split in the overthrust, for, east of the lowland, north-dipping Cambrian quartzite forms two detached hills, separated by pre-Cambrian rocks. The southern hill is a nearly direct continuation of the Thomas Hill trend; the northern hill, Mt. Pleasure, is a similar monoclinal block, repeated by faulting. The overthrust apparently splits, and the interpretation given in the section in Figure 3-B shows it as forming an imbricate slice.

Another low hill similarly capped by north-dipping Chickies quartzite, southeast of Mt. Pleasure on the border of the Phoenixville quadrangle, is another downfold of this quartzite. Its north edge is overlapped by Triassic rocks, which are believed to lie on limestone and, therefore, to cover the overthrust between the quartzite and the limestone. The fault is interpreted as following the edge of the lowland, which is believed to be underlain by the limestone.

The pre-Cambrian core of the Welsh Mountain uplifted mass is cut off abruptly on its south side by a relatively straight east-west fault, which is called the Brandywine Manor fault (Figs. 2 and 4). The downthrow on this break is to the south and it was, therefore, at first interpreted to be a normal tension fault, and was so described in the report on the New Holland quadrangle.[4]

Tension faulting would normally take place after the period of compression had ended, probably in Triassic time. This fault has not been traced in the pre-Cambrian rocks east of Font, where there is a remnant of down-faulted Chickies quartzite. It does not seem to affect the Triassic rocks on the northern border of the pre-Cambrian, and evidence of Triassic age is lacking. An oval-shaped area of anorthosite at the south edge of the pre-Cambrian rocks of the Welsh Mountain anticline is bordered by a narrow band of quartz monzonite, as mapped by Bascom (Fig. 2). This intrusive mass is believed to have been the deepest-lying rock in the area and to have been brought to the surface at the center of greatest uplift of the anticline. The oval shape of the mass is probably the original shape of the intrusion. If the Brandywine Manor fault is a normal tension fault, the structure in the down-dropped block to the south should be similar to that in the upraised block. The rocks adjacent to the fault in the block to the south, however, indicate a syncline opposite the core of the Welsh Mountain anticline. Chickies quartzite and, in places, Harpers phyllite and Vintage dolomite were undoubtedly inclosed in a syncline between the Welsh Mountain and the Barren Hills

[4] Anna I. Jonas and George W. Stose: *op. cit.*, p. 19.

anticlines, and this syncline was cut off on its north side by the Brandywine Manor fault. Such relations cannot be accounted for by a normal tension fault, but suggest an overthrust on a high-angle fault of the Welsh Mountain anticline, southward over the intervening syncline. The Brandywine Manor fault, and the quartzites in the syncline south of it, bow gently around the massive anorthosite mass in a manner that also suggests that these rocks were subjected to compression during, or after, faulting. Nowhere is there coarse brecciation of the rocks adjacent to the fault, as is common with normal tension faults of Triassic age in the region, but the rocks near the fault show the effects of shearing, such as mylonitization of the anorthosite and schistose quartzite, indicating faulting under compression.

The available evidence in the distribution of, and structure in, the formations on the two sides of the break indicate, therefore, that the rocks on the north side were overthrust southward on a high-angle plane, onto the rocks to the south. There is no doubt that there was relatively great upward movement of the pre-Cambrian rocks north of the fault with respect to the Cambrian quartzites and limestones south of the fault. Such relative movement is not only shown at the anorthosite body at the center of uplift but also south of the west end of Welsh Mountain, where pre-Cambrian rocks are adjacent to, and stand higher than, Ledger dolomite on the south, indicating a vertical displacement of at least 2000 feet (Fig. 2). Although the Brandywine Manor break continues along the south side of the Dry Hill anticline, west of Welsh Mountain, the throw on this part of the fault is minor, and the main displacement terminates abruptly at the west end of Welsh Mountain (Fig. 4). Here, the Elverson overthrust on the north side of the anticline swings south around the end of the mountain, cutting across the strike of the Cambrian quartzites into the pre-Cambrian rocks, and joins the Brandywine Manor fault. The Elverson fault evidently is here a tear fault. The mass affected by the great upward movement is clearly limited to the Welsh Mountain mass between these two faults. It seems reasonable to conclude, therefore, that the Welsh Mountain pre-Cambrian core was thrust upward on a steep or nearly vertical break on its south side and that the upthrust block tore loose at its west end.

The fault contact in most places is not exposed, and the structures in the rock at the fault are seen in few places. One of the few places where the fault contact is exposed is at Cornog, on the East Branch of Brandywine Creek (Fig. 5). Here, the anorthosite just north of the fault has numerous shear zones with layers of mylonite, which dip 60° South. The quartzite south of the fault is, likewise, sheared to a sericite-quartz schist, also dipping 60° South, and a quartz vein and spring are at the

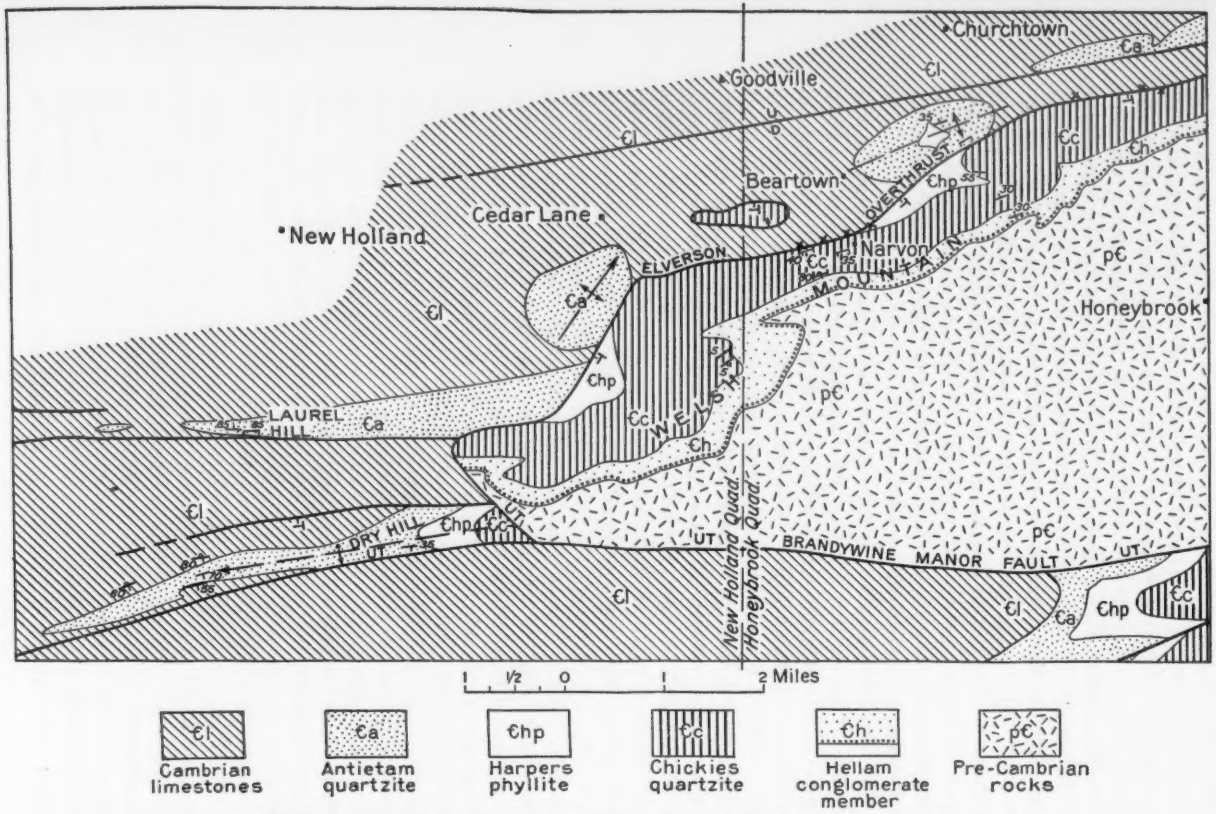


FIGURE 4.—*Geological map of the western end of the Welsh Mountain anticline*

The Elverson overthrust unites with the Brandywine Manor upthrust at the west end of Welsh Mountain, where the upthrust mass tore loose from the two elongated anticlines of the west-plunging end of the fold. (T) Overthrust side of low-angle fault, (UT) upthrust side of high-angle compression fault. Numerous iron ore banks (X) are shown along the Elverson overthrust.

contact. On the upland near Brandywine Manor, large masses of residual vein quartz mark the fault. A road cut here exposes highly sheared sericitized quartzite with the schistosity dipping 45° South, accompanied by a quartz vein. The shear planes of the quartzite are pencilled by grooves that dip S 45° E. North of Brandamore Station, on the Wil-

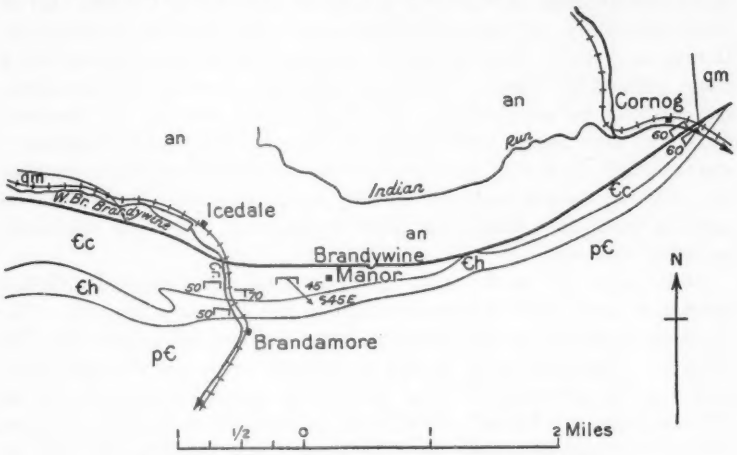


FIGURE 5.—*Map of the vicinity of Brandywine Manor*

Showing rock structures adjacent to the Brandywine Manor fault. Schistosity in quartzite, and mylonite in anorthosite, dip 45° to 60° South. Grooves on schistose planes dip S 45° E. (€c) Chickies quartzite, (€h) Hellam conglomerate member, (an) anorthosite, (qm) quartz monzonite, (pC) undifferentiated pre-Cambrian rocks.

mington division of the Pennsylvania Railroad, along the West Branch of Brandywine Creek, a bed of quartzite, which dips 70° South, shows minute folds overturned toward the northwest. The quartzite is greatly sheared, the schistosity dipping 30° to 50° South (Fig. 5).

The 45° to 60° South dip of the schistosity in the quartzite and the adjacent anorthosite does not accord with the postulated upthrust of the Welsh Mountain mass on a vertical or steeply inclined fault plane, but rather favors the interpretation that the rocks to the south were overthrust on those to the north, the grooving indicating the horizontal direction of movement to be N 45° W. Although overthrusting of younger rocks on older, as this would be, is not unknown in the Appalachians, it is exceptional. The existing relations, if produced by overthrusting in that direction, could result only from the overthrusting of a syncline enclosing Cambrian rocks on a core of pre-Cambrian rocks in an anticline, and would represent considerable horizontal displacement. If this were so, the intervening syncline and the southern limb of the Welsh Mountain

FIGURE 4.—*Geological map of the western end of the Welsh Mountain anticline*

The Elverson overthrust unites with the Brandywine Manor upthrust at the west end of Welsh Mountain, where the upthrust mass tore loose from the two elongated anticlines of the west-plunging end of the fold. (T) Overthrust side of low-angle fault, (UT) upthrust side of high-angle compression fault. Numerous iron ore banks (X) are shown along the Elverson overthrust.

anticline should appear along the fault to the east and west, where the movement on the fault decreased away from the point of maximum overthrusting. Instead, the syncline of Cambrian quartzite in the intervening syncline south of the fault rises eastward and is lost in the pre-Cambrian, except for the small remnant of the quartzite at Font. Westward the major displacement on this fault terminates abruptly at the west end of Welsh Mountain, without any indication of these postulated structures. It may be possible that the Welsh Mountain mass was upthrust on a plane inclined 70° South, beneath the load of quartzite and limestone, and produced the south-dipping schistosity. Or it may be that the fault plane was more nearly vertical and has been tilted by later compression and rotation. It is even possible that the south-dipping schistosity near the fault and the grooving on the schist planes represent a forward movement of the Barren Hills mass after the uplift of the Welsh Mountain anticline was complete.

At the west end of the main upfaulted mass, an elongated anticline extends 4 miles west of the mountain in the limestone lowland. The Antietam quartzite in this anticline forms the low hill called Dry Hill (Fig. 4). The anticline is faulted on both its north and its south sides, and, near the mountain, Chickies quartzite is enclosed between the faults. This anticline is believed to have been a minor fold at the west-plunging end of the Welsh Mountain anticline when it first began to rise, and the fault on its south side was a continuation of the Brandywine Manor fault in the early stage of uplift, before the main body of the anticline tore loose along the tear fault. The Laurel Hill anticline to the north, a similar elongated anticline, is believed to have been another minor fold at the plunging end of the Welsh Mountain anticline, its eastern end having been overridden by the Welsh Mountain mass as it moved forward on the Elverson overthrust. Both these minor folds are tightly compressed, so that the beds stand nearly vertical, which accounts for their elongated character and straight outlines. The northern fold is especially tightly compressed, the observed bedding being vertical and the schistosity steep to vertical, and the bounding faults nearly straight. These folds, therefore, appear to be due to sharp upfolding and upthrusting on steep fault planes of the resistant Cambrian quartzite, structure similar to the Mill Lane anticline, described later.

From the evidence presented the writer favors the hypothesis that the Brandywine Manor fault is an upthrust of the Welsh Mountain anticline on a vertical or steeply inclined plane, possibly modified by later less intense overthrusting of the Barren Hills anticline from the south. According to this interpretation the steps in the uplift were as follows: The Welsh Mountain anticline began as an upthrust of pre-Cambrian

rocks, pushed by a force acting in an upward, and northwestward, direction, the pressure apparently centering at the anorthosite mass (Fig. 6). This force was not caused by the intrusion of the anorthosite while molten,

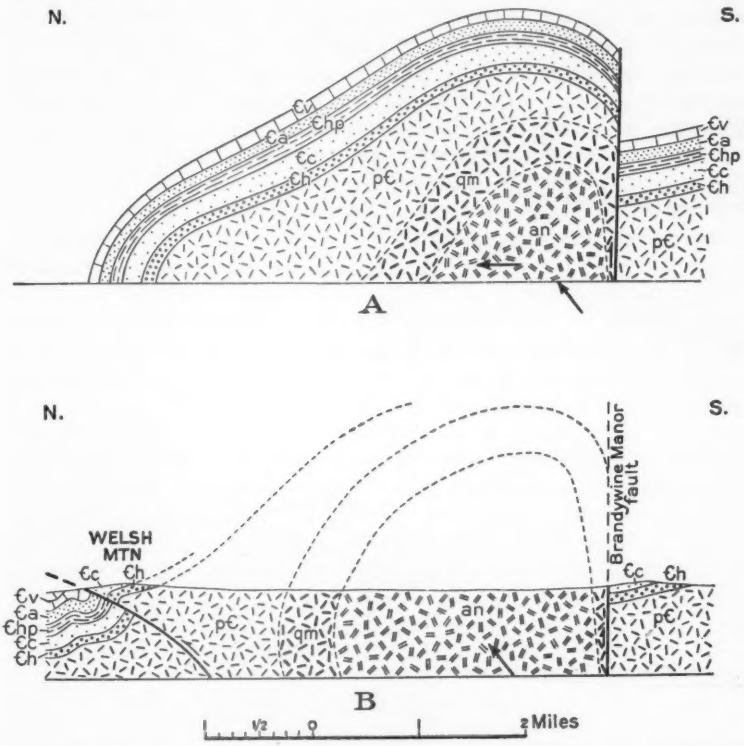


FIGURE 6.—*Structure sections showing hypothetical stages of uplift of the Welsh Mountain anticline*

(A) Hypothetical cross-section of the Welsh Mountain anticline in early stage of uplift. Showing upthrust of anorthosite mass on Brandywine Manor fault and forward movement of the uplift mass, resulting from a diagonal upthrust from below. (B) Cross-section of the anticline after further upthrust and overthrust. This structure was probably modified by later forward movement of the Barren Hills block, at the south, producing the southeast-dipping schistosity in the quartzite and anorthosite, Cv, Vintage dolomite; Ca, Antietam quartzite; Chp, Harpers phyllite; Cc, Chickies quartzite; Ch, Hellam conglomerate member; an, anorthosite; qm, quartz monzonite; pC, undifferentiated pre-Cambrian rocks.

as that occurred much earlier, in pre-Cambrian time, but this resistant crystalline mass seems to have transmitted the pressure from below, possibly because of its dense compact homogeneous character and great

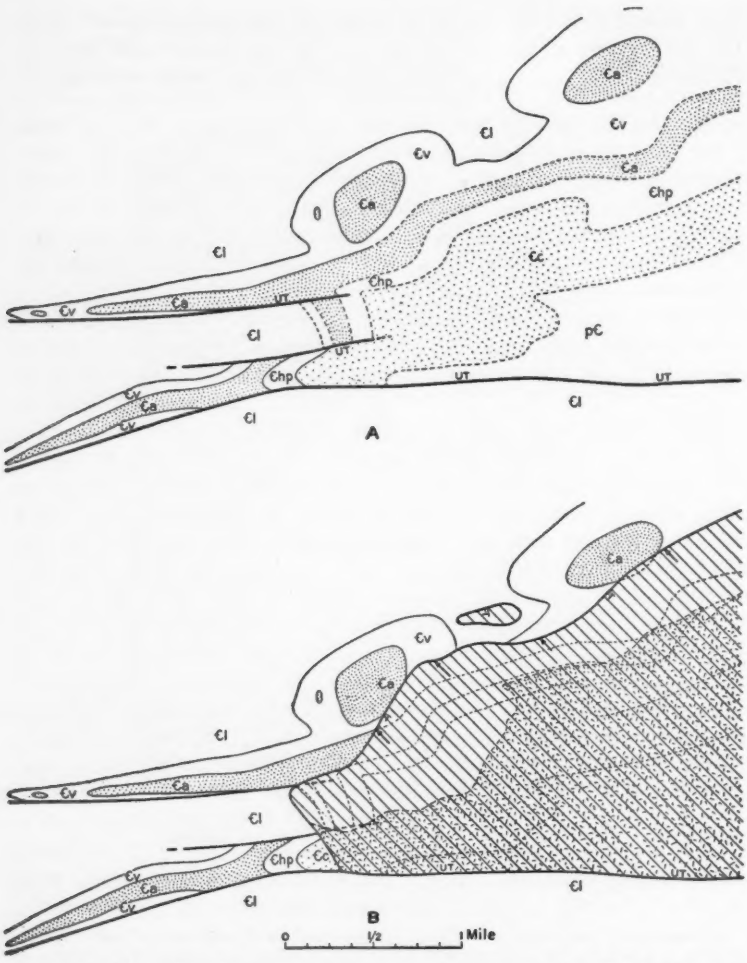


FIGURE 7.—*Maps showing hypothetical stages of uplift of the Welsh Mountain anticline*

(A) Early stage of uplift. Showing major uplift along a steep break along the south margin of the mass and minor breaks on the flanks of westward-plunging anticlines. (B) Buried structures of the early stage, overridden by the Elverson overthrust, as the upthrust mass rose along the steep Brandywine Manor fault at the south and tore loose at its west end. Shaded area, overthrust and upthrust block; hachured area, pre-Cambrian rocks of upthrust block; Cl, younger Cambrian limestones; Cv, Vintage dolomite; Ca, Antietam quartzite; Chp, Harpers phyllite; Cc, Chickies quartzite; pC, pre-Cambrian rocks; T, overthrust on low-angle fault; Arrows, direction of movement; UT, upthrust on high-angle plane.

depth, or possibly because it directly overlay a deeper channel up which other molten rock was seeking to rise. An east-west nearly vertical break in the overlying rocks allowed the pre-Cambrian rocks to rise nearly vertically, thus relieving the vertical component of the compression, as shown diagrammatically in Figure 6-A. This break became the Brandywine Manor fault. The residual horizontal component of the force pushed the uplifted rocks northward and folded them into an anticline. In its early stages, this anticline was gentle and low and plunged steeply westward from the center of uplift, but two minor attenuated anticlines of Cambrian quartzite extended far out into the limestone west of the main uplift. In this early stage the Brandywine Manor break extended along the south side of the southern of these anticlines (Fig. 7-A). As the uplift increased, the horizontal component of the uplifting force pushed the anticline forward until it broke at the front, and the uplifting Cambrian quartzites on the north limb of the anticline were thrust northward on the Elverson overthrust. The earlier structures of the anticline were overridden by the forward movement of the overthrust mass, as shown diagrammatically in Figure 7-B. In this process the greatly upthrust mass was torn loose at its west end, and the Elverson overthrust joined the Brandywine Manor fault at the west end of the mountain as a great tear fault. In this description the term normal tension fault is used for a normal fault, one that is vertical or steep and hades to the downthrow, which shows no evidence of having taken place under compression. Upthrust is used for a normal fault that is vertical or steep and shows evidence of having taken place under compression. Tear fault is used for a nearly vertical break that cuts directly across the structure; the displacement may be either horizontal or vertical; in this instance, the displacement is largely vertical.

BARREN HILLS ANTICLINE

This fold, although it occupies a depression between the higher Welsh Mountain and the Mine Ridge anticlines, is, nevertheless, an anticline, for the Cambrian quartzites, which form the Barren Hills at the westward-plunging end of the fold, are locally preserved in synclines on both flanks of the fold toward the east. A long narrow band of north-dipping Cambrian quartzite in the syncline between the Barren Hills and the Welsh Mountain anticlines is enclosed to, and beyond, Brandywine Manor, and an outlying area of this quartzite in the syncline is preserved as far east as Font (Fig. 2). The quartzite in the syncline is elsewhere cut out at the surface by faulting. The triangular area of Vintage dolomite 3 miles south of Honeybrook is probably a remnant of the deeper part of this syncline in a fault block. The south side of this dolomite

area is cut off by quartzites in a minor plunging fold of the Barren Hills anticline, which apparently are overthrust from the south. This may represent the overthrusting on the north side of the Barren Hills anticline, previously referred to in the description of the Welsh Mountain anticline. As there stated, the quartzites near Brandywine Manor and Cornog have south-dipping schistosity, which apparently was produced by northward overthrusting of the Barren Hills anticline, which is believed to have taken place after the upthrusting of the Welsh Mountain mass on the Brandywine Manor fault. The complex faulting north and west of the Barren Hills suggests that the rocks in the intervening syncline were crushed between the upthrust pre-Cambrian rocks of the Welsh Mountain anticline and those of the overthrust Barren Hills anticline.

The syncline on the south limb of the Barren Hills anticline is represented by south-dipping Cambrian quartzite near Wagontown and north and northeast of Downingtown. Between these two places, it is cut out by a fault, which is interpreted as an overthrust on the north side of the Mine Ridge anticline, and is called the Gap overthrust.

Minor westward-plunging folds in the Cambrian quartzite of the Barren Hills are reflected in the sinuous edge of the hills and the winding course of the narrow band of Kinzers formation in the limestone valley (Fig. 2). The two northern minor anticlines are broken on their south side by faults with downthrow to the south. This displacement might have been produced by an upthrust from the north, produced by the drag from the larger upthrust in this direction on the Brandywine Manor fault. A minor fold on the south side of the anticline has been broken on its north side by a curving fault, which is interpreted as a northward overthrust, probably induced by the drag of the Gap overthrust from the south, where the movement on that fault was greatest (Fig. 9). At Limeville the Chickies quartzite in this minor fold is overthrust on Ledger dolomite, and the fault is called the Limeville overthrust.

The pre-Cambrian core of the Barren Hills anticline, which is 2½ to 3 miles wide at the west, expands eastward to a width of 6 miles, and is overlapped by the Triassic southwest of Phoenixville. The Cambrian quartzites on the south limb of the fold make a continuous band from a point northwest of Downingtown to Devault. North of Bacton the quartzite ridge is offset by a normal tension fault trending east-northeast, with downthrow to the south (Fig. 8).

At Devault the pre-Cambrian rocks in the core of the fold are brought in contact with gently folded Elbrook limestone and Ledger dolomite in the valley to the south. This might be the result of another east-west normal tension fault with downthrow to the south, but the fault trace appears to be a curved line around the limestone lowland (Fig. 8) and

not a series of broken straight faults, such as most normal faults in the Appalachian region are. The attitude of the quartzites at their termination on either side of Devault indicates that they were locally bowed

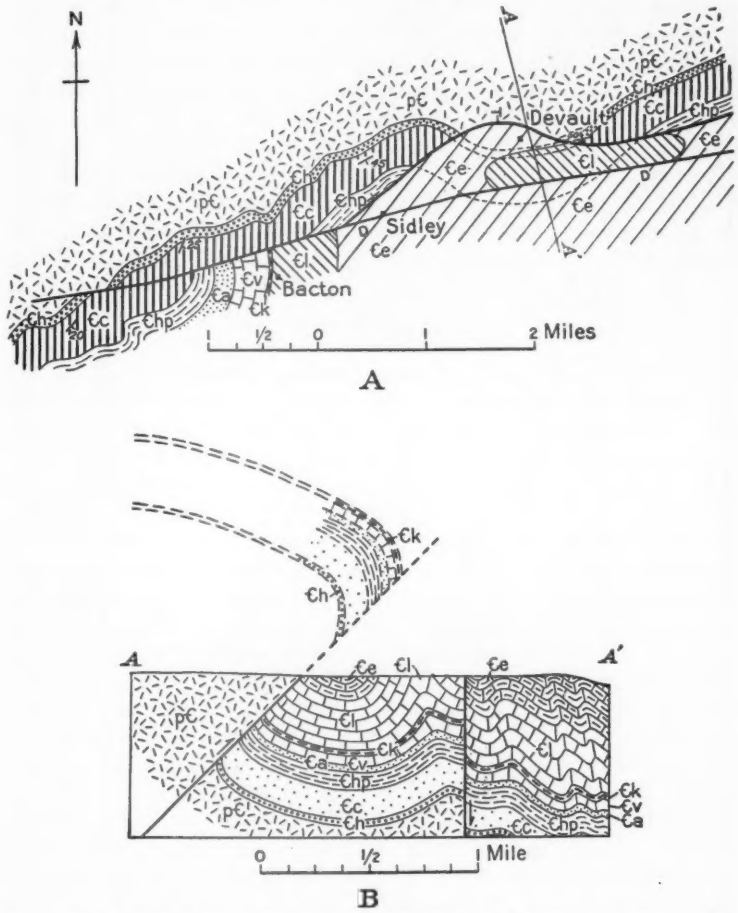


FIGURE 8.—*Geological map and section of the south edge of the Barren Hills anticline at Devault*

(A) Showing the Devault overthrust and the Sidley normal fault. Note overturned beds at Devault. (B) Cross-section through Devault, showing theoretical interpretation of the Devault overthrust. Ce, Elbrook limestone; Cl, Ledger dolomite; Ck, Kinzers formation; Cv, Vintage dolomite; Ca, Antietam quartzite; Chp, Harpers phyllite; Cc, Chickies quartzite; Ch, Hellam conglomerate member; pC, pre-Cambrian rocks; T, overthrust side of low-angle fault; D, downthrow side of normal tension fault.

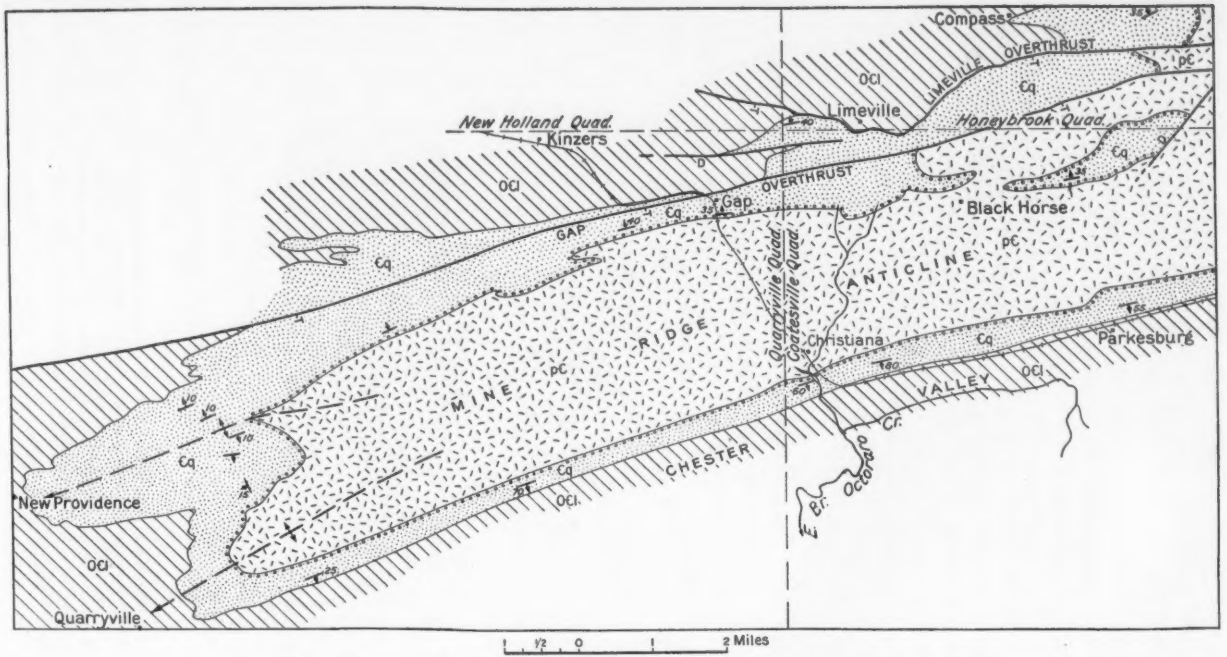


FIGURE 9.—*Generalized geological map of the western part of the Mine Ridge anticline*

Showing Gap and Limeville overthrusts and Chester Valley syncline. OCl, Cambrian and Ordovician limestones; Cq, Cambrian quartzites and phyllite; pC, pre-Cambrian rocks; T, overthrust side of low-angle fault; D, downthrow side of normal tension fault.

up into a gentle anticline with north-south axis, which was truncated by the fault. The limestones south of the fault show no evidence of this anticline, as they should if the limestones are down-faulted, but, instead, the axes of the gentle folds trend east-west. The Hellam conglomerate, which normally dips south off the Barren Hills anticline, is locally overturned just east of Devault and dips 50° north. From the data available, it is concluded that the rocks in a minor anticline on the limb of the main Barren Hills fold are overturned toward the south and are broken by a local overthrust, the pre-Cambrian rocks in the core of the fold being overthrust southward onto the limestone (Fig. 8-B). This is similar to the faulting at Elverson and east of Thomas Hill, where the Elverson overthrust cuts across minor wrinkles on the sole of the thrust on the north limb of the Welsh Mountain anticline, except that the thrusting is in the reverse direction.

MINE RIDGE ANTICLINE

The Mine Ridge anticline is best shown and most clearly defined south of Gap, in the Quarryville quadrangle.[5] There the pre-Cambrian core, 2½ miles wide, is flanked on both sides by Cambrian quartzites. The north-dipping quartzites on the north limb form Mine Ridge, and south-dipping quartzites on the south limb form the North Valley Hills (Figs. 2 and 9). The quartzites on the south limb are much thinner and are less resistant than those on the north limb, and the North Valley Hills are, therefore, lower and less prominent than Mine Ridge. Northeast of Quarryville the fold divides into two west-plunging anticlines, and the Cambrian quartzites wrap around the pitching ends of these folds as they dip beneath the limestone of the valley (Fig. 9).

The north side of the anticline is broken by a thrust fault, and at the north foot of Mine Ridge, at Gap, the upper quartzites are cut out, and the Chickies is overthrust on Vintage dolomite. This fault is called the Gap overthrust. Three miles east of Gap the quartzites are entirely cut out by this fault, and pre-Cambrian rocks are thrust over the Cambrian quartzites of the Barren Hills anticline. East of this point the Cambrian quartzites do not again appear on the north flank of the fold, and the anticline gradually narrows to a width of 1 mile north of Downingtown (Fig. 2). Three miles northeast of Downingtown the anticline plunges eastward, and the quartzites wrap around the ends of minor pitching folds, but the structure is complicated by several faults. The Gap overthrust, however, continues along the north side of the anticline to a point north of Exton, where it is terminated by a diagonal normal fault (Fig 10-A). This fault is probably a later tension fault, for it is marked by numerous

[5] E. B. Knopf and A. I. Jonas: *loc. cit.*

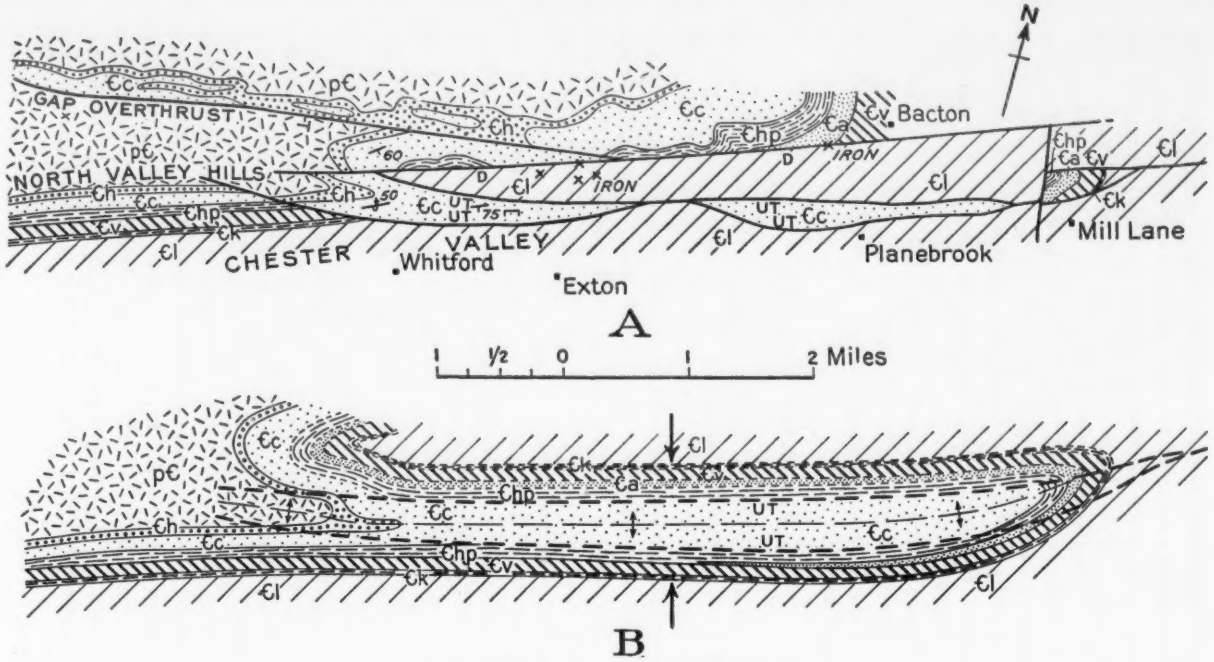


FIGURE 10.—*Maps of the Mill Lane anticline*

(A) Geological map of the up-faulted Mill Lane anticline. The Mine Ridge anticline, bordered on its north side by the Gap overthrust, splits into two anticlines at its eastward-plunging end and is largely terminated by a nearly east-west normal fault. The southern anticlinal arm is attenuated into an elongated fault block, called the Mill Lane anticline. (B) Plan of a theoretical restoration of the Mill Lane anticline before faulting. €l, Ledger dolomite; €k, Kinzers formation; €v, Vintage dolomite; €a, Antietam quartzite; €hp, Harpers phyllite and Antietam quartzite in part; €c, Chickies quartzite; €h, Hellam conglomerate member; pC, pre-Cambrian rocks; UT, upthrust side of high-angle fault; D, downthrow side of normal tension fault.

iron ore pits, ferruginous geodiferous cherts, and springs, and cuts diagonally across the structures. At the west the Chickies quartzite is faulted against the Ledger dolomite in the valley to the south, and farther east the Harpers, Antietam, Vintage, Kinzers, and Ledger formations are consecutively faulted against the Ledger dolomite to the south.

Northeast of Whitford (Fig. 10-A) the southern minor fold on the plunging end of the Mine Ridge anticline takes the form of an elongated

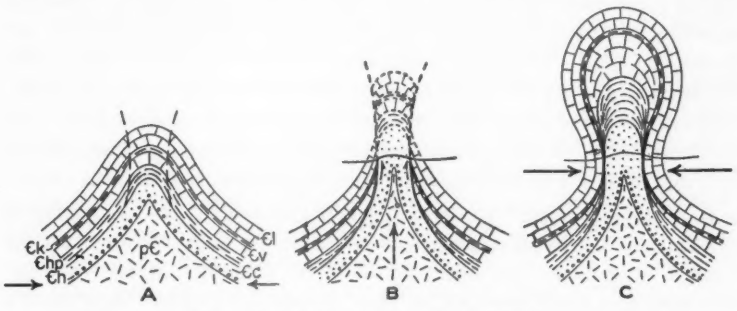


FIGURE 11.—*Theoretical interpretations of the Mill Lane structure*

(A and B) Cross-sections of two stages of development. Showing the Chickies quartzite tightly compressed and up-faulted on both sides and sheared through the overlying beds to the Ledger dolomite. (C) The same result if produced by squeezing out of beds on the limbs of the folds (not supported by evidence). Cl, Ledger dolomite; Ck, Kinzers formation; Cv, Vintage dolomite; Ca, Antietam quartzite; Chp, Harpers phyllite and Antietam quartzite; Cc, Chickies quartzite; Ch, Hellam conglomerate member; pC, pre-Cambrian rocks.

fault block of Chickies quartzite, bounded by nearly parallel faults, which extends eastward nearly continuously for 5 miles and terminates in the small quartzite knoll north of Mill Lane. The anticlinal character of this fold, which is referred to as the Mill Lane anticline, is shown at its east-plunging end north of Mill Lane, where the Harpers phyllite, Antietam quartzite, and Vintage dolomite in normal sequence above the Chickies quartzite, are preserved between nearly parallel faults (Fig. 10-A). The Chickies quartzite in the narrow block is bordered on both sides by Ledger dolomite, and the four normally intervening formations are faulted out. The only place in this fault block where the quartzite can be examined in outcrop is in an old quarry, 1 mile northeast of Whitford. It is here sheared to a quartz-sericite schist, which crumbles readily and has been quarried or dug for sand and white sericitic clay. The layering, which is in part bedding, as shown by pebbly beds, dips 75° south. Thick homogeneous quartzose beds are mashed to a quartz schist with white sericitic clay bands, and the closely spaced cleavage planes, which also dip 75° South, are coated with sericite and have scattered tourmaline

crystals on them. The schistosity is evidence of intense compression and differential movement.

The structure is interpreted as a greatly compressed anticline, bounded on both sides by steeply dipping fault planes, along which the Chickies quartzite has been up-thrust nearly vertically, shearing through the overlying beds, and now protrudes through the limestone (Figs. 11-A and B). This relation might have resulted from an isoclinal anticline so tightly compressed laterally that the beds between the massive Chickies quartzite and the thick-bedded Ledger dolomite were squeezed out on the flanks and forced upward into the crest of the fold (Fig. 11-C). The character of the top of the fold is portrayed by the remnant of its crest, preserved at its plunging end, in the fault block north of Mill Lane. As no thickening of the formations at the crest is there shown, the interpretation suggested in Figure 11-C is not supported by the evidence.

In a previous paper [6] the writer included this structure with examples of similarly faulted elongated narrow anticlines of Antietam quartzite flanked by younger limestone, northeast of Lancaster, which were interpreted as a horst produced by later downfaulting of the limestone on steep normal tension faults on the sides of the anticline. This tension faulting was attributed to Triassic movement, for which there was some warrant, as down-faulted blocks, or grabens, enclosing Triassic rocks, and therefore of later age, are found less than 3 miles to the north of the so-called anticlinal horst near Lancaster. In the Mill Lane anticline the evidence of intense compression, such as the schistose character of the quartzite, the extensive development of sericite on the numerous cleavage planes, and overturned dips of the quartzite in places, strongly favors the view that in the Mill Lane anticline the whole movement took place under compression, while the anticline was rising. It is even possible that all the so-called anticlinal horsts described by the writer in the previous paper [7] are upthrusts of tightly compressed anticlines on high-angle fault planes and due to compression and not to later down-faulting of the limestone on the flanks of the folds. The series of so-called anticlinal horsts described in that paper form a nearly continuous chain from the anticlines on the east-plunging end of the Chickies-Chestnut Hill uplift, northwest of Lancaster, to those on the west-plunging end of the Welsh Mountain uplift, and seem to be related structures, and may be comparable to the Mill Lane anticline.

[6] George W. Stose: *New type of structure in the Appalachians*, Geol. Soc. Am., Bull., vol. 35 (1924) p. 475-479.

[7] *Ibid.*

LATER CROSS FAULTS

West of Downingtown the Chester Valley syncline and the Mine Ridge anticline are cut by a north-south cross fault along which the outcropping formations are offset (Fig. 2). The movement on the fault appears to be largely horizontal, the block on the east side having moved south half a mile with respect to that on the west, and it is called a shear fault. Not only are the formations offset in this direction but so also is the Gap overthrust. The movement on this fault is, therefore, later than the general deformation of the region and may be of late Triassic age. However, a Triassic dike in west Downingtown is not noticeably offset.

A similar nearly parallel cross fault cuts the rocks west of Mill Lane, but the general effect appears to have resulted largely from vertical, rather than horizontal, movement. This cross fault cuts the Mill Lane anticline at right angles, and the part east of the fault has been relatively dropped, as it shows the crest of the Mill Lane fold (Fig. 10-A). The steep parallel faults that bound the anticline, however, are offset in a manner that does not harmonize with a simple vertical movement. The fault appears to terminate in the east-west normal tension fault south of Bacton, and the limestone formations on opposite sides of the fault are not offset in the same direction (Fig. 2). The movement was, therefore, probably complex.

AGE OF THE STRUCTURES

The structures described are known to have been produced later than early Ordovician time because the Conestoga limestone, which is probably of Beekmantown age, is involved in the structure. The rocks were folded, broken, and overthrust during a period of great compression and mountain-making, and most of the structures, therefore, are believed to have been formed during the Appalachian revolution at the close of Carboniferous time, when most of the mountain-building of the Appalachians occurred. Earlier in Paleozoic time, there was minor upfolding, such as the incipient rising of the Mine Ridge anticline, indicated by the thinning of the early Cambrian sediments on its crest and in the Chester Valley basin to the south.[8] Also, this area was gently uplifted in early Ordovician (pre-Conestoga) time, and some of the rocks then at the surface were eroded, so that, when the Conestoga limestone was deposited, it overlapped the eroded edges of formations down to the Ledger dolomite, and, in the Coatesville region, down to the Harpers phyllite (Fig. 2). The transverse faults at Downingtown and at Mill Lane are of

[8] George W. Stose and Anna I. Jonas: *Ordovician overlap in the Piedmont Province of Pennsylvania and Maryland*, Geol. Soc. Am., Bull., vol. 34 (1923) p. 521-524.

somewhat later date and may be of late Triassic age. Possibly, also, some of the east-west normal tension faults south of Devault are of late Triassic age, because similar nearly parallel normal faults in the vicinity of King of Prussia, 5 miles east of the Phoenixville quadrangle, affect the Triassic sedimentary rocks and are, therefore, in part at least, of later age.

U. S. Geological Survey, Washington, D. C.
Manuscript received by the Secretary of the Society, February 9, 1937.
Read before the Geological Society, December 27, 1936.
Published with the permission of the Director of the U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 48, PP. 1001-1078, 8 PLS., 12 FIGS. JULY 1, 1937

CAMBRIAN AND ORDOVICIAN OF NORTHWESTERN VERMONT

BY CHARLES SCHUCHERT

CONTENTS

Page

Introduction 1003
General statement 1003
Previous work 1005
The historic "Georgia Group" 1011
The strange breccias and the intraformational conglomerates 1014
Breccias 1014
Intraformational conglomerates 1016
Thrust contacts 1016
Times of orogeny 1018
Orogeny in Vermont 1018
Geosynclines and geanticlines 1018
Orogeny in southern Quebec 1020
Lower Cambrian 1022
Formations and faunas 1022
General statement 1022
Monkton quartzite 1022
Winooski dolomite and marble 1022
Contact between Winooski and Mallett 1023
Mallett dolomite 1024
Break between Mallett and Parker 1025
Parker slate 1025
Long erosion interval 1026
Significance of the dissimilar Cambrian faunas in northwestern Vermont and in the Taconic Mountains 1026
Detailed description of localities 1028
Northeast of Highgate Springs 1028
East of Highgate Springs 1030
Two miles north 15° west of Highgate Center 1031
Southeast of Swanton village 1031
Donaldson farm, east of Swanton 1034
Three miles southeast of Swanton 1034
Near Fonda quarry, south of Swanton Junction 1035
North of St. Albans Bay 1036
South of St. Albans Bay village 1037
Conner farm, southwest of St. Albans 1037
Near the mouth of Rugg Brook, east of St. Albans Bay 1038
Burlington 1038
Jones Hill 1038
Taconic sequence, West Castleton 1038

Page

Middle Cambrian 1039
General statement 1039
Formations and faunas 1039
Rugg Brook conglomerate 1039
St. Albans slate 1040
Detailed description of localities 1041
Adams pasture, west of St. Albans 1041
Two and a half miles north of St. Albans 1042
Rockledge 1042
Conner farm, southwest of St. Albans 1043
Near the mouth of Rugg Brook 1043
Mill River 1044
Upper Cambrian 1044
General statement 1044
Formations and faunas 1046
Milton dolomite 1046
Mill River limestone breccia 1047
Hungerford slate 1047
Rockledge limestone breccia 1049
Georgia slate 1052
Gorge formation 1052
"Hungerford" slate northwest of Highgate Center 1055
"Milton" dolomite 1057
Detailed description of localities 1057
Cobble Hill 1057
Allen Brook near Cobble Hill 1058
Milton 1059
Two miles south of Georgia Plains 1059
Georgia Center and west 1059
Four miles southwest of St. Albans 1061
Less than three miles south of St. Albans 1062
Mayre ledge, two miles south of St. Albans 1062
Adams pasture, St. Albans 1062
Three and a half miles north of St. Albans 1063
Rockledge 1064
Four and a half miles north of St. Albans 1065
Six miles north of St. Albans 1065
Six and a half miles north of St. Albans 1066
Highgate gorge 1067
Ordovician 1070
Formations and faunas 1070
General statement 1070
Highgate formation 1070
Post-Highgate emergence and orogeny 1073
Corliss breccia 1074
Grandge slate 1075
Detailed description of localities 1076
Highgate Falls, Missisquoi River 1076
Highgate Center 1077
One and one-fourth miles northeast of Highgate Center 1078

Page
Two miles northeast of Highgate Center 1078
Two and one-third miles northwest of Highgate Center 1078
Southeast of Burlington 1078

ILLUSTRATIONS

Figure Page
1. Geographic and geologic sketch map of northwestern Vermont 1004
2. Published sections in northwestern Vermont 1006
3. "Georgia Group" as interpreted by Walcott 1012
4. Theoretic sections across northern New England 1019
5. The celebrated Georgia section 1024
6. Sequence from Russell quarry west 1031
7. Sequence from Swanton marble quarry to Donaldson farm 1032
8. Erosion contact between Parker slate and Mallett dolomite 1034
9. Sequence north of Swanton Junction station 1035
10. *Bovicornellum vermontense* Howell 1040
11. The Russell road-metal quarry 1056
12. Sequence from Donaldson farm east to Highgate road 1066

Plate Facing page
1. Rockledge breccia 1016
2. Upper Cambrian of Vermont 1017
3. Champlain thrust near Burlington, Vermont 1022
4. Lower and Upper Cambrian of Vermont 1023
5. Vermont bioherms 1048
6. Vermont breccias 1049
7. Lower Ordovician of Vermont 1070
8. Lower Ordovician contacts in Vermont 1071

INTRODUCTION

GENERAL STATEMENT

No Cambrian area in all eastern North America is more interesting stratigraphically, or shows more complicated structure, than that of northwestern Vermont and the adjacent portion of New York. Even after nearly a century of endeavor, it cannot be said that the entire sequence of the formations is now known, and it will not be satisfactorily adjusted until the succession of the formations in the Eastern Sequence of northwestern Vermont is also unraveled. The present paper is restricted to the Cambrian and Lower Ordovician stratigraphy and faunas of the Central Sequence of the northwestern part of Vermont (Fig. 1), as originally defined by Keith. The Upper Cambrian and Ordovician stratigraphy as now presented varies considerably from that given by the writer in 1933.[1]

[1] Charles Schuchert: *Cambrian and Ordovician stratigraphy of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 25 (1933) p. 353-381.

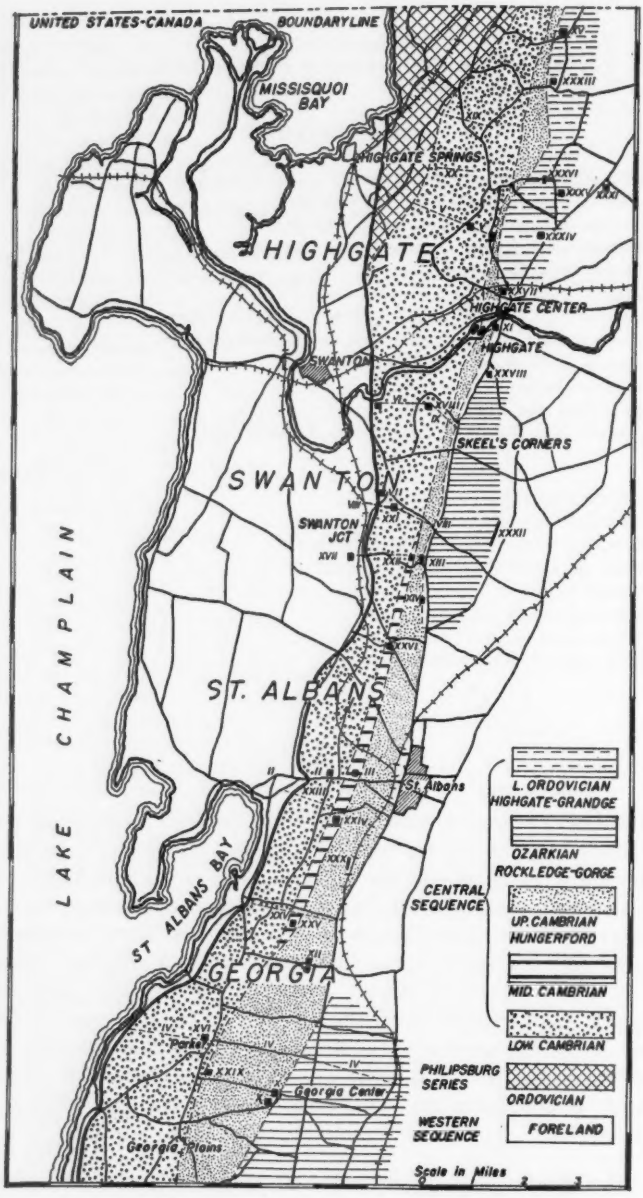


FIGURE 1.—*Geographic and geologic sketch map of northwestern Vermont*

Localities for fossils, and described sequences, shown by black squares and thin broken lines. Roads in solid thick lines. Champlain fault (Logan's Line) indicated by thickest line at western side of shaded area. Shaded area represents the Central Sequence. Based on United States Geological Survey maps of St. Albans and Milton quadrangles, which show topography and much more geographic detail.

PREVIOUS WORK

The rise of modern geology in New York and Vermont may be said to have begun at least one hundred years ago,[2] with the second edition of Professor Amos Eaton's *Geological Text-book*, published at Albany in 1832. Eaton, graduate of Williams, student under the elder Silliman at Yale at the age of forty, lawyer, surveyor, botanist, chemist, geologist, was "an original genius of profound and far-reaching intellect," and has often been called the "Father of American Geology." Out of his "Primitive Series" came the "Granular Quartz" (1818), which was the first formation of the Cambrian system in America to have a name, and which was rechristened the Cheshire quartzite by Emerson in 1892. Eaton's term "Argillite" (1820-1839) embraced the Cambrian and Ordovician slates of New York, and also those of Vermont, a part of which, beginning with 1861, came to be known as the Georgia Group of Edward Hitchcock.

The first to describe the local geology of the Taconic Mountains was Professor Chester Dewey, who presented in 1820 the results of field work in the area about Williamstown, Massachusetts, and thence westward to Troy, New York, in the Hudson Valley.

The Taconic system, that great irritant of unacknowledged ignorance, was defined in 1842 by another of the teachers of geology at Williams—Professor Ebenezer Emmons. Emmons held that all of this system, which he at first believed to be devoid of fossils, was older than the Potsdam sandstone of the New York system, and lay "beneath the zone of animal life." The several state geologists of New York, and chiefly Hall, insisted, on the other hand, that there could be no Paleozoic formations older than the Potsdam, as the latter rests on ancient granite. In fact, all the leading geologists of the day who were familiar with the region regarded the Taconic system as of the same age as the Hudson River group, which was thought to have the Potsdam sandstone at its base; and in all the area of the Mississippi Valley east to the Appalachian Mountains, no sedimentary formation was then thought to be older than the Potsdam. The proposal of the Taconic system by Emmons in 1842 accordingly precipitated the most acrimonious and long-drawn-out discussion in the annals of American geology, the so called "Taconic controversy."

The state geologist of Vermont, Charles B. Adams, "a conchologist of great merit," who had the support of Professor Edward Hitchcock, voiced the prevailing view in 1846, when he said that the Taconic system is "supposed by some geologists to be only a part of the New York system altered by the action of heat. Others again regard them [the Taconic formation] as a distinct and older system," i.e., older than the Champlain

[2] For a more extended account of this history, with references, *see* C. D. Walcott: *Correlation papers; Cambrian*, U. S. Geol. Surv., Bull. 81 (1891) p. 91-122.

FIGURE 2.—*Published sections in northwestern Vermont*

(A) Edward Hitchcock's section (1861) eastward across Snake Mountain. (*a-f*) Ordovician formations of Western Sequence; (*k*) talus; (*g*) Red Sandrock, now known to be the overthrust Monkton formation of Lower Cambrian age; (*mm*) supposed fault. The great overthrust beneath *g* was not seen.

(B) Edward Hitchcock's section (1861) of the "Georgia formation" from Lake Champlain east through St. Albans and Aldis Hill (= Fig. 1, traverses II and III). (A and B) Ordovician of Western Sequence. The following formations are of the overthrust Central Sequence: (C) Lower Cambrian Winooski and Mallett, and (N) Parker slate; (D) Middle Cambrian slate, overlain by (E) Mill River limestone breccia; (S and F) Hungerford slate of the Upper Cambrian. (G, H, K, and L) are of another overthrust mass of Proterozoic age, a part of the Eastern Sequence, all included by Hitchcock in the Georgia formation.

(C) Keith's diagram (1933) from St. Albans Bay to Aldis Hill, showing five overthrusts (= Fig. 1, traverses II and III). Eastern Sequence: 1, 2, 3, and 8, undifferentiated pre-Cambrian and ? older Paleozoic. Central Sequence: Lower Cambrian, (4) Winooski and Mallett dolomites, and (5) Parker slate. Middle Cambrian, (6) St. Albans slate. Upper Cambrian, (7) Mill River breccia, and (8-9) Hungerford slate. (10) Western Sequence.

division of the New York system, or what is now known as the Ordovician.[3]

The "Red Sandrock" of Snake Mountain, Vermont, one of the bones

[3] C. B. Adams: *Second annual report on the geology of the State of Vermont* (1846) p. 104.

of contention in the Taconic controversy, had been recognized by Adams as early as 1847, but the strata to which the term was applied had their first clear description by Edward Hitchcock in his *Geology of Vermont* (1861) (Fig. 2, A). In 1847, Rev. Zadock Thompson and Rev. S. R. Hall, members of the Vermont Geological Survey, found "fossil shells and fragments of trilobites" east of Highgate Springs (not at Snake Mountain), in strata which Hitchcock in 1861 called Red Sandrock, and which are now divided into the Monkton, Winooski, and Mallett formations of the Lower Cambrian. In the Vermont report of 1847, James Hall says that the trilobite fragments "are all of the central portion of the buckler," and are like those of "*Conocephalus* and *Olenus*." These bucklers, Hall thought, were "most nearly like one in the Clinton group" of New York. This trilobite was later described by Hall, and it is now known as *Ptychoparella adamsi*, a form quite unlike anything found in the Clinton formation. Hall, however, was cautious, and added that he could learn little of these fragmentary fossils, and not enough of their occurrence to infer safely their "real geological position". But Adams adds that he also has two brachiopods from this same Red Sandrock, and that they "resemble *Atrypa hemispherica* of the Clinton group"—an impossible identification. Inasmuch as Hall grouped the Clinton with the Medina sandstone, and as Hitchcock thought that he had also seen *Arthrophycus harlani* in the Sandrock, and in view of the further (but erroneous) supposition that the Red Sandrock rests without exception "upon the Hudson River group" of the Middle Ordovician, all the geologists of that time were led to correlate the Red Sandrock of northwestern Vermont with the Medina sandstone at the base of the Silurian. Even the accomplished W. B. Rogers was certain in 1851 that the Red Sandrock was of Medina age, and he had the backing of Emmons, Adams, and Logan. Not until 1861 did anyone see a fault—it turned out to be a big overthrust—beneath the Sandrock, and then Emmons placed one in the midst of the Ordovician (Fig. 2, A). Overthrusting was unknown in those days, and the dread of faulting led Hitchcock to say in 1861:

> "The theorist's desire for faults . . . is no more readily satiated than the miser's desire for gold—the more he obtains the more he wants. One instinctively shrinks back from a theory involving numerous faults." [4]

Eventually, it was the recognition of this marked overthrust, first clearly noted by Sir William Logan in 1863, that was to lead to the solution of the Taconic riddle. The controversy continued to rage because all the geologists who took part in the discussion (1) relied mainly upon lithology in correlation, (2) believed that superposition is always dependable in indicating the succession, and (3) either misunderstood or badly

[4] Edward Hitchcock *et al.: Report on the geology of Vermont* (1861) vol. 1, p. 375.

interpreted what fossils were then known. Even as late as 1886, cleavage was mistaken for bedding in Vermont, and the strata were said to dip steeply to the east throughout the 5-mile width of the Georgia Group, whereas, in reality, they lie in broad undulations.

The unfolding of the Taconic controversy would require a volume. The interested reader should consult George P. Merrill's readable chapter, "The Taconic question".[5] Another good account, especially in its application to Vermont, is that by Edward Hitchcock.[6]

It seems curious now that fragments of *"Ptychoparia adamsi,"* a Lower Cambrian fossil, should have been used to prove the Silurian (Medinan) age of what is now called the Monkton quartzite. The first entire trilobites in the Taconic system were found as early as 1844 in the black slates at Reynolds Inn, north of Bald Mountain, Washington County, New York, by an entomologist and physician, Asa Fitch, of Salem, New York, and were described in that same year by Emmons as *Atops trilineatus* and *Elliptocephala asaphoides*. At that time, these fossils in themselves, unsupported by the evidence of correctly interpreted superposition, could not prove that the rocks were older than the Potsdam, or that they were of Lower Silurian (*Olenus*) age, as Hall believed; hence their discovery only aggravated the controversy. This was especially true of *Atops trilineatus*, a trilobite without eyes, which Hall wrongly maintained was the same as *Triarthrus becki* of the Ordovician, which has eyes in the normal position.

Some years afterward, a Vermont farmer, Noah Parker, found entire trilobites in his road-metal quarry (Fig. 1, loc. XVI) and gave them to Zadock Thompson, who, in turn, presented them to James Hall. The latter described them in 1859 as *Olenus* (now *Olenellus*) *thompsoni*, *Olenus* (now *Mesonacis*) *vermontanus*, and *Peltura* (now *Bathynotus*) *holopyga*, and regarded them as indicative of the Hudson River group; in other words, he wrongly referred to the Middle Ordovician the rocks yielding these unmistakable Lower Cambrian fossils. As John M. Clarke says in his interesting book on the life of James Hall:

> "The error was now an egregious one for in the interval the character of the primordial fauna had been clearly demonstrated and Hall himself, in his correspondence with Barrande and others, had conceded the primordial type of just such trilobites."[7]

The two geologists who were the first to begin to unravel the structure of northwestern Vermont, to give a more correct interpretation of the strata, and thus to bring some daylight into the Taconic problem, were

[5] G. P. Merrill: *The first one hundred years of American geology* (1924) p. 594-614. Yale University Press, New Haven.

[6] Edward Hitchcock: *op. cit.*, p. 434-447.

[7] J. M. Clarke: *James Hall of Albany* (1921) p. 389.

Canadians. The leader in this work was the remarkably gifted director of the Geological Survey of Canada, Sir William Logan, and his helpmate was Elkanah Billings, a lawyer by profession but a born paleontologist. The former, between 1861 and 1863, followed the structural trends from Canada south through Vermont to Burlington; noted in a clear manner in his *Geology of Canada* (1863) the major north-south faulting, the east-west overthrusting of the Red Sandrock composed of "Primordial" formations, and the superposition of the latter on fossiliferous Ordovician; and finally, in 1865, presented a geological map. His associate described the new forms in 1861, making use of those published earlier by Hall, and then sent his printed results to Joachim Barrande, of Bohemia. Barrande was already familiar with some of the Primordial fossils of the United States and Canada, and had presented a paper about them in 1860 before the Geological Society of France. Agreeing with Emmons that the Taconic system was older than any of the New York system, Barrande said that the fossils of Parker's farm in Georgia township were of the age of his "First or Primordial fauna", and much older than those of the Olenus zone, which was correlated with the Potsdam. These fossils are in reality older than Barrande's. However, it was Barrande's publications that confirmed the work of Logan and Billings and established the fact that a number of the fossils of the Taconic system were older than any known in the New York system. These results of Barrande and Billings must have warmed the heart of the exiled Emmons, who died in North Carolina late in 1863.

Twenty years after Logan's book appeared, Walcott began his remarkable and highly successful search for fossils in the Cambrian strata of the Taconic Mountains of eastern New York and western Vermont. His epochal paper of 1886 [8] redescribed Hitchcock's Red Sandrock formation and his Georgia Group. Two years later,[9] he contributed a paper on the Taconic system of Emmons, and in this and in a number of later papers recorded the many places at which Cambrian fossils had been found, thus laying the foundation for all studies of the subsequent historical geology of the region. His results helped greatly toward the adjustment of the Taconic problem, but the true sequence of the Cambrian epochs was not determined until the field season of 1888, when Walcott established the succession on Manuel's Brook, Conception Bay, southeast Newfoundland, where it was learned for the first time in the history of stratigraphy that the rocks containing the *Olenellus* fauna conformably underlie the

[8] C. D. Walcott: *Second contribution to the studies on the Cambrian faunas of North America*, U. S. Geol. Surv., Bull. 30 (1886) p. 13-20.

[9] C. D. Walcott: *The Taconic system of Emmons*, Am. Jour. Sci., 3d ser., vol. 35 (1888) p. 229-242, 307-327, 394-401.

Primordial zone, which Barrande had placed at the base of the Paleozoic, because it rested on granite.[10]

Although great progress had been made between 1860 and 1865 along paleontological and stratigraphical lines, the significance of Logan's observations on the structure of northwestern Vermont was not fully recognized. The faulting, which thrusts older formations over the Ordovician all along the St. Lawrence River front in Quebec and in northwest Vermont, was called by Logan the "great fault" and the "break or overlap"; it was described by him in 1863 and illustrated in five sections in the Atlas of 1865.[11]

In 1884, S. W. Ford [12] pointed out the overthrust nature of the Lower Cambrian in the Hudson River series, which, at Schodack Landing, New York, rests on the Middle Ordovician (Normanskill). James D. Dana [13] noted these faults in his *Manual*, and in the fourth edition called them "great flexures, great faults, and overthrust flexures".

Even with these published observations regarding the structure, Vermont geologists continued to start their field work in the west, with the nearly horizontal and highly fossiliferous formations exposed along the shore of Lake Champlain, and to proceed eastward across the strike and structure, hoping to trace the unaltered fossiliferous western formations into the more or less altered eastern supposed equivalents. Such a method of correlation, overlooking or failing to understand the profound overthrust relations, which negate the superposition, could lead to no final conclusion; and it remained for Arthur Keith to bring to full appreciation the work begun so many years ago by Logan.

Keith began his mapping of the geology of the Taconic Mountains in southern Vermont and eastern New York in 1909, and four years later extended his work into northwestern Vermont. In 1921, he wanted paleontological help, and invited the writer to accompany him. In the course of a motor trip from Rutland to Highgate Falls, the writer became so favorably impressed with Keith's methods and views that he insisted on the immediate publication of his results, and a paper presenting these appeared in the *American Journal of Science* in February 1923.[14] Keith's work, joined with that of Logan and Billings and Walcott, has furnished the key that will unravel the detail of the complicated geology west of the Green Mountains, for the truth can be arrived at only by detailed stratigraphic work, based on carefully collected fossils, and checked by a correct

[10] C. D. Walcott: *Correlation papers; Cambrian*, U. S. Geol. Surv., Bull. 81 (1891) p. 113.

[11] W. E. Logan: *Geology of Canada* (1863) p. 283-285, 820, 851, 855, 857; *Atlas* (1865) map IV.

[12] S. W. Ford: *Note on the discovery of primordial fossils in the town of Stuyvesant, Columbia County, New York*, Am. Jour. Sci., 3d ser., vol. 28 (1884) p. 35-37.

[13] J. D. Dana: *Manual of geology*, 4th ed. (1895) p. 386, 528.

[14] Arthur Keith: *Cambrian succession of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 97-139.

interpretation of the exceptional structure. By pursuing these methods, Keith came to see the actual succession of the formations in the Champlain synclinorium. There are, between Lake Champlain and the Green Mountains, he says, "from three to seven faults in any single cross section, and the width of the belt between the faults is only from three to ten miles." [15] The formations can be understood only by arranging them in "three sequences, originally far apart east and west . . . now jammed together along great thrust faults." [16] The slightly disturbed formations of the foreland, which lie west of the Champlain fault, form the Western Sequence, and to the east of this lie the Central Sequence and the Eastern Sequence.

Since 1921, the writer has made thirteen trips to northwestern Vermont, spending anywhere from a few days to a month there in most years, and being accompanied at different times by Keith, Sayles, Prindle, Raymond, Howell, Clark, Dunbar, Longwell, and Adolph and Eleanora Knopf, with all of whom the stratigraphic succession and the fossils have been discussed. Percy E. Raymond has described in another paper the Upper Cambrian and Lower Ordovician trilobites and ostracodes, and B. F. Howell, who is conversant with the Vermont Cambrian, has contributed a paper on the Middle Cambrian fauna, both of which will be published in a forthcoming number of the BULLETIN.

THE HISTORIC "GEORGIA GROUP"

As early as 1847, State Geologist C. B. Adams and Rev. Zadock Thompson had examined the Georgia and St. Albans areas in which are the slates that later were named the Georgia group, and which figured much in the Taconic discussion. The term, Georgia group, was proposed by Edward Hitchcock for strata that include "two (perhaps three) terrains in Vermont," and that were erroneously estimated to be more than 10,000 feet thick. The group "is identified by . . . organic remains" found on Noah Parker's farm. Hitchcock goes on to say that Emmons believed the Georgia slate to be "the uppermost member of the Taconic system".

Hitchcock [17] presents a diagram of the area eastward from St. Albans Bay (Fig. 2, B) with the following sequence: At the west are "Hudson River shales, followed by dove-colored limestone of the Upper Hudson River group". These are now correlated with the Middle Ordovician and are of the Western Sequence. Over these the diagram shows in conformable (now known to be thrust) relation the Red Sandrock (= Winooski and Mallett), followed by the Georgia slate, with the horizon of *Olenellus thompsoni*. A limestone breccia (= Mill River) is noted. East of the

[15] *Op. cit.*, p. 104.
[16] *Op. cit.*, p. 103.
[17] Edward Hitchcock: *Geology of Vermont*, vol. 1 (1861) p. 357-386.

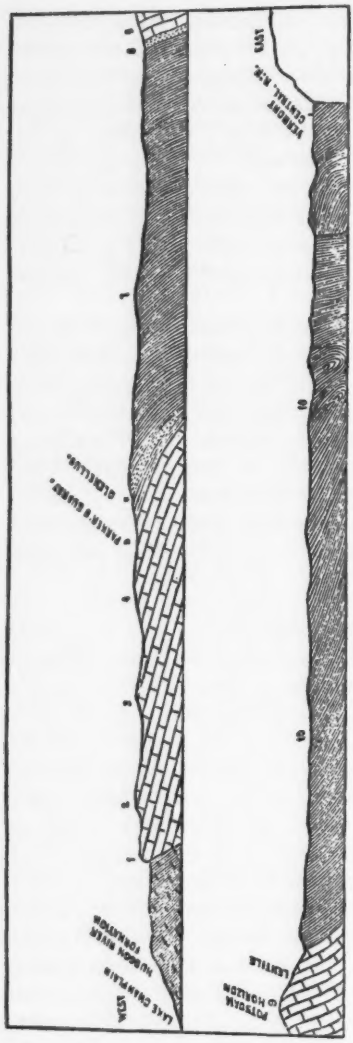


FIGURE 3.—*"Georgia Group" as interpreted by Walcott*

Beds 1-10 are described in detail on pages 1013 and 1014. This sequence extends from Lake Champlain east for more than 4 miles (Fig. 1, traverse IV). At the east are overthrust Proterozoic formations. Compare with Figure 5.

city of St. Albans, much more "Georgia" slate is indicated, followed by "brecciated quartz rock" at Aldis Hill. The ages of the last two formations are unknown.

Walcott, beginning his classic discussion of the Georgia terrane in 1886, says:

> "As we now know the Georgia Formation, it appears that Dr. Emmons was correct in placing it below the Potsdam sandstone, as was also done by Mr. Billings and later writers. The typical Georgia Formation, as developed in the town of Georgia, Franklin County, Vermont, consists, as seen at the base, of a great thickness of magnesian limestones that pass, in their upper portions, into an arenaceous magnesian limestone that is overlaid by a belt of arenaceo-argillaceous shales, and this by a great thickness of a purer argillaceous shale that, high up, carries a brecciated limestone conglomerate."[18]

Walcott's figure illustrating the Georgia sequence is here reproduced (Fig. 3), as it helps much in understanding the following description. Originally thought to be all of Lower Cambrian age (hence Walcott's period term, Georgian[19]), the section is now known to include much Upper Cambrian and probably some Ordovician. Walcott thought that the Georgia series in its type area was approximately 10,000 feet in thick-

[18] C. D. Walcott: *Second contribution to the studies on the Cambrian faunas of North America*, U. S. Geol. Surv., Bull. 30 (1886) p. 14-19.

[19] C. D. Walcott: *Correlation papers; Cambrian*, U. S. Geol. Surv., Bull. 81 (1891) p. 360.

ness, but recent work shows that it has about 2000 feet of Lower Cambrian, more than 1000 feet of Upper Cambrian, and probably some Ordovician.

A summary of Walcott's section, with its modern equivalents, follows:

	Thickness (Feet)
LOWER CAMBRIAN (Figs. 1, traverse IV; 3, 5)	
Beds 1–5 of Walcott = Winooski-Mallett dolomites (1000 feet).	
1. Blue-gray, massive dolomites	35
2. Dark buff and bluish black dolomite with angular fragments of a blue-gray dolomite	200
3. Gray, massive dolomite more or less mottled	475
4. Reddish pink dolomite	100
5. Gray, arenaceous, massive dolomite	190
Bed 6. "Georgia shales" of Walcott = Parker slate (Fig. 1, loc. XVI). Dark-blue micaceous shales, with numerous fossils, found at various places on the Parker farm. A little road-metal quarry at the base of the western hill, about 125 feet above the Mallett dolomite, yielded the fossils described by Hall. Strike, "N. 30° E, dip 8° to 12° E." Walcott gives the thickness as 200 feet, but it is nearer	300
In the upper third of the "Georgia shale", on the western hill, are found erosion remnants of two blue-gray dolomites weathering a rusty red or salmon color, and beneath these is another zone for fossils. (For lists of fossils, see page 1026.)	
Bed 6a. Parker dolomite. Above the Parker slate is a capping of blue-gray dolomite that makes the top of Parkers Cobble and the top of the Lower Cambrian. It dips about 20° E. Walcott interprets this as "a lenticular mass left by erosion, and resting conformably on the shales beneath, each dipping 10° eastward." This dolomite, however, is not a lentil but a remnant of a widespread dolomite zone to the south (where it thickens to 50 feet or more) and north of the Parker farm, and lying upon the Parker slate.	
Disconformity and erosional break. All Middle Cambrian absent until a place 2.5 miles to the northeast is reached.	
BASAL UPPER CAMBRIAN	
Bed 7 of Walcott = Hungerford slate. "East of the Parker quarry the rocks are argillaceous shales with occasional layers of hard gray limestone, one-half of an inch to two inches thick, that carry numerous fragments of a linguloid shell. Strike of shales near top of (7) N. 40° to 60° E., dip 60° SE. . . . 3500 feet." This dip, however, is of cleavage and not of bedding (Pl. 4, fig. 2). The actual thickness of these Upper Cambrian (Hungerford) slates may not be more than	700
A marked break occurs at the top of Bed 7, and it is followed unconformably by Beds 8 and 9 (= Rockledge breccia). Walcott regards Beds 1-7 inclusive as of the more typical Georgia Terrane, and notes "a decided change in the fauna" beginning with Beds 8 and 9. (For further detail, see pages 1047-1049.)	

	Thickness (Feet)
UPPER CAMBRIAN	
Beds 8 and 9 of Walcott = Rockledge breccia.	
8. "Light-gray quartzite . . . 50 feet." Actual thickness probably not more than	10
9. "Gray limestone in massive layers. . . . Many of the beds of limestone appear to have been broken up into fragments and recemented in situ. Average strike of limestone beds N.50°E., dip 50°-90°; average dip 60° SE. . . . 1700 feet." This dip is again that of cleavage. Elsewhere, Walcott calls Bed 9 "the great lenticular mass". What he refers to is the formation now called the Rockledge breccia, which covers a wide expanse of land less than a mile west of Georgia Center, and has a probable thickness of	30–40
The fauna, Walcott says, "is Cambrian in character, and . . . approaches that of the Upper Cambrian or Potsdam sandstone." Concerning other fossils supposed to have been found in this breccia, Ulrich[20] says that they are now known to have been collected by Walcott in the Lower Cambrian of Washington County, New York. Ulrich and Bassler[21] redescribed these phyllopods in 1931. This, then, clears up an erroneous record that has long puzzled the writer.	
(For further description, see pages 1049-1052.)	
No break in deposition.	
Bed 10 of Walcott = "Georgia slate" of Keith 1932.	
Walcott describes his Bed 10 (Fig. 3) as follows: "Argillaceous shales, very similar to those in the Parker ledge, continue on up to the opposite side of the line of the Vermont Central Railroad track. At the base the shales rest conformably against the limestone of 9, and above appear to be cut off by a fault. Strike N.50°E., dip 60° to 80° SE." The thickness, according to Walcott, is 3500-4000 feet. Keith places it at about	2000
As no fossils are known from this higher "Georgia slate", all that can be said about its age is that its lower part at least is of latest Cambrian time, for it follows without depositional break upon the Rockledge breccia. It may be that the higher part is of Lower Ordovician age (= Grandge slate).	

THE STRANGE BRECCIAS AND THE INTRAFORMATIONAL CONGLOMERATES

Breccias.—The most difficult type of formation to interpret and to correlate in northwestern Vermont is the limestone breccia, of local origin, and presumably made by a transgressing sea. There are four of these basal breccias. One is the Rugg Brook, composed of dolomite and underlying the Middle Cambrian. The other three are made up of limestone pieces and blocks that look much alike; one is the Mill River, at the base of the Hungerford slate; another is the Rockledge, which lies beneath the Georgia formation (Pl. 1; Pl. 6, fig. 1); both are of Upper

[20] E. O. Ulrich: *Personal communication.*

[21] E. O. Ulrich and R. S. Bassler: *Cambrian bivalved Crustacea of the order Conchostraca,* U. S. Nat. Mus., Pr., vol. 78 (1931) art. 4.

Cambrian time. The third limestone breccia is the Corliss, at the base of the Grandge slate of the Lower Ordovician.

In addition to these depositional basal breccias, the area has two thrust breccias, which are not strikingly different in appearance from the type just described. One of these lies at the base of the Hungerford slate, and has been confused with the Mill River depositional breccia, because of its place in the sequence. The other is beneath the Highgate formation (Pl. 6, fig. 2). These thrust breccias have more large blocks than the other type and include more kinds of sedimentary rocks, which are also in more angular pieces.

The Rockledge breccia (Pl. 1; Pl. 6, fig. 1) at the base of the "Georgia" formation, and, to a less extent, the Mill River breccia beneath the Hungerford slate, have great blocks of limestone measuring from 20 to more than 120 feet in length; these blocks were local reefs made by algae, which grew faster than the adjacent mud bottoms were built. They were common during Hungerford time, and other similar ones developed on the mud bottoms of the Lower Cambrian. During the intermediate land times, these reef limestones tended to become residuals, and the invading seas either covered them with muds or pounded them to pieces, making the Mill River, Rockledge, and Corliss basal breccias. Where the invading seas encountered protruding lenses of dolomite, the latter were broken up into the breccia known as the Rugg Brook.

The Lower Ordovician Quebec series of eastern Canada is likewise replete with local breccias, but these recur many times in the same formation and never appear to be basal to a transgressive deposit. Such breccias have been interpreted by Bailey, Collet, and Field [22] as the result of local slipping and breaking up of reef limestones along the sea bottom, caused by earthquakes and faulting in a tectonic belt. The basal breccias of Vermont, however, are local marine accumulations, formed at or near the places where the reef limestones grew, or to which they gravitated during land intervals. The encroaching seas either pounded to pieces those reefs that were encountered in shallow water, or piled around the lentils in deeper water the thin-bedded limestones picked up elsewhere out of the Gorge or Highgate formations. In other words, the land over which these seas transgressed was uneven, with low erosion remnants of horizontal limestones or dolomites, and here and there a reef lentil. Where there were no ridges of harder material, but only a substratum of shale, the Lower Ordovician (Highgate) limestones rest on the Upper Cambrian Gorge slate, the Upper Cambrian muds (Hungerford) lie on the Middle Cambrian shale (St. Albans), and the Middle Cambrian shale overlies the Lower Cambrian shale.

[22] E. B. Bailey, L. W. Collet, and R. M. Field: *Paleozoic submarine landslips near Quebec City*, Jour. Geol., vol. 36 (1928) p. 577-614.

Intraformational Conglomerates.—Northwestern Vermont is remarkable also for its many intraformational conglomerates, which either make entire layers or occur as scattered pieces in the dolomites. Where they are in beds, these aggregations of angular slabby pieces of limestone or dolomite occur most commonly in the thin-bedded strata of the Upper Cambrian (Gorge dolomite) and more rarely in the Lower Ordovician Highgate limestone. In the upper division of the Gorge formation, at least eight such breccia beds appear in quick succession (Pl. 2). Elsewhere, in thin-bedded dolomites, such thin slabby pieces are seen scattered in the Milton formation and in the lower part of the Gorge formation, both of the Upper Cambrian. In the Lower Cambrian, the intraformational conglomerates are much rarer, but they are present in the dolomites of the Winooski formation and more commonly in the Mallett dolomite. None is known in the Middle Cambrian slates, nor in any of the other slate formations.

The depositional evidence suggests accumulation in shallow water, where the thin limy beds were subject to seasonal exposure and sun-cracking and were then submerged and the fragments washed together by the storm-swept waters. There may also have been occasional warping, creeping, and slipping of the sea bottom, caused by earthquakes and faulting. Such disturbances may have given rise to local ridges of thin-bedded dolomites and limestones, which were later torn to pieces by the shallow-water waves, which washed the fragments about and in places piled them up to make the intraformational conglomerates.

THRUST CONTACTS

Exposures in various places show the different members of the Lower Cambrian of the Central Sequence resting upon various formations of the Ordovician of the Western Sequence. To the south of Burlington and at Snake Mountain, it is the oldest Cambrian formation—the Monkton—that is thrust upon the Ordovician; farther north, the overriding beds are generally different parts of the Winooski dolomite, the most instructive accessible contact being at Lone Rock Point near Burlington; at Fonda quarries and in southern Quebec, it is the Mallett dolomite that plays this rôle. It is apparent, therefore, that the known basement of the overthrust Central Sequence is everywhere made up of highly competent formations belonging to the Lower Cambrian.

For more than a mile to the north of St. Albans Bay village, there is a grand display of the marbleized Ordovician limestones of the Highgate Springs sequence, with the Winooski dolomites overthrust upon them. The contact is nowhere actually exposed, but in places the two series can be seen within a few feet of one another. The Winooski shows no alteration, nor is there a basal mylonite, but the marbleized limestone

ROCKLEDGE BRECCIA

Showing blocks of bioherm limestone, 10 to 15 feet long, surrounded by smaller pieces, some of which have trilobites of the same age as those found in the limestone *in situ*. West of Georgia Center (Fig. 1, loc. X). Upper photograph by A. C. Swinnerton; lower, by C. O. Dunbar.

FIGURE 1. UPPER GORGE THIN-BEDDED DOLOMITES AND LIMESTONES
Highgate gorge (Fig. 1, loc. XI). R. W. Sayles is standing on a 3-foot bed of intraformational conglomerate. The cliff has several limestones rich in trilobites. Higher in the cliff is the thrust breccia at the base of the Highgate formation (see Pl. 6, fig. 2). Photograph by A. C. Swinnerton.

FIGURE 2. UPPER GORGE INTRAFORMATIONAL CONGLOMERATE
Close view of the conglomerate shown in figure 1. Eight similar, but thinner, beds, are exposed in this cliff. Photograph by A. C. Swinnerton.

beneath is much altered, with here and there a sandy shale zone that is mashed and rolled. This long exposure shows, furthermore, that the north-south thrust plane is not a strictly horizontal one, but that it is undulatory, with depressions as much as 60 feet deep; the dip of the strata toward the east is less than 10 degrees.

Contacts like those described in the previous paragraph are also seen for nearly 2 miles to the south of St. Albans Bay village. An actual contact between the Winooski dolomite and the marble of the Highgate Springs series is exposed to the west of the Everitt school. Here, and to the north as well, lower Winooski forms the basement of the thrust sheet. The plane of the thrust dips 20°E., and, although the Ordovician limestones are changed to marble, no breccia or mylonite is present.

The best place to examine the St. Lawrence thrust is in the Fonda quarry, a little southwest of Swanton Junction (Fig. 1, loc. XVII). Logan [23] showed that the Highgate Springs series, consisting of Trenton, Black River, and probably Chazy formations, is "in an inverted attitude" in this quarry, and dips eastward, with the greatest overturn at the east. On the northern slope the Mallett lies thrust from east to west. The actual overriding contact upon the limestone is best seen as one goes from the north around to the western side of the quarry hill. The marbleized limestone is decidedly metamorphosed and schistose, whereas the overriding Mallett dolomite is only fractured and faulted. Going around to the south side of the quarry hill, one has a view of the entire high north wall of the quarry.

The most instructive of the thrust contacts is to be seen along the Lake Champlain shore at Lone Rock Point, north of Burlington (Pl. 3); the cliff-making Winooski dolomite is thrust over much-crumpled Canajoharie slate of the Trenton series (with graptolites). The Winooski projects considerably, and its under side is fluted and striated, showing that the thrust was from the southeast.

Another fine exposure, of a totally different nature, is about 1.5 miles north of St. Armand in southern Quebec, on and near the Grove Hill farm of W. C. Brown. At this locality, it is the Mallett dolomite that has moved across the Lower Ordovician of the Philipsburg block; the magnesian limestones of the latter are brecciated into a mylonite and are scattered between large blocks, which have been moved about and offset in small faults. One can walk over square rods of the Lower Ordovician, where the overriding Mallett has been eroded away, and can note the effect of the overthrusting. In places, the thrust plane is at an angle of 35 degrees or more, but on the average the angle is about 10 degrees.

Still another exposure of a thrust plane—the best for detail—is seen

[23] W. N. Logan: *Geology of Canada* (1863) p. 280-281.

on the Donaldson farm, northwest of Skeels Corners (Fig. 1, traverse IX), where a long thrust breccia at the base of the Upper Cambrian (Hungerford) overrides the Lower Cambrian (Parker). This contact is described on page 1067.

A different thrust breccia is exposed in the Highgate gorge (Pl. 6, fig. 2), where the earliest Ordovician (Highgate) overrides the Gorge formation of the Upper Cambrian (p. 1077).

TIMES OF OROGENY

Orogeny in Vermont.—The earliest Paleozoic orogenic event in the region with which this paper is concerned took place during Lower Ordovician time, at the close of Highgate deposition. It is known as the Quebec phase of the Taconian disturbance, and has been observed only in an area of one square mile. It is described further in connection with the Highgate formation.

To the west of the Green Mountains in Vermont, no strata now present are known to be younger than the highest Middle Ordovician (Canajoharie shale and Highgate Springs series), and all the beds east of the Western Sequence are involved in the thrusting and folding of the region. Farther north, in the area of the St. Lawrence plain, in southern Quebec, there is, however, a thick series of younger Ordovician age (Lorraine), and, southeast of Montreal and near Three Rivers, the younger strata of this system (Richmondian) are seen. Younger formations are not present in this general region, but, on the island of St. Helen's in the St. Lawrence River opposite Montreal, a dike of volcanic agglomerate, in strata that are known to be of late Middle Ordovician age, yields late Lower Devonian and early Middle Devonian fossils. In the eastern part of the St. Lawrence geosyncline, to the east of the overthrust Green Mountains, on Lake Memphremagog, southern Quebec, fossiliferous Middle Silurian and late Lower Devonian formations lie in unmistakable angular unconformity upon folded Ordovician strata. The younger formations are found in synclines, apparently made by the Acadian orogeny of late Devonian time. Accordingly, Vermont, west of the New Brunswick geanticline, was subjected to intense folding and thrusting, first at the close of the Ordovician (Taconian movement), and again in late Devonian time (Acadian movement), when the Acadian and New England regions as a whole suffered a renewal of intense orogeny.[24]

Geosynclines and Geanticlines.—In eastern Vermont and in New Hamp-

[24] For further discussion of these orogenic events, *see* Arthur Keith: *Further discoveries in the Taconic Mountains*, Geol. Soc. Am., Bull., vol. 24 (1913) p. 680.

Charles Schuchert: *Orogenic times of the northern Appalachians*, Geol. Soc. Am., Bull., vol. 41 (1930) p. 701-724.

Charles Schuchert and C. R. Longwell: *Paleozoic deformations of the Hudson Valley region*, Am. Jour. Sci., 5th ser., vol. 23 (1932) p. 305-326.

shire lay the ancient New Brunswick geanticline, which separated the St. Lawrence geosyncline from the Acadian trough still farther east (Fig. 4). The original St. Lawrence geosyncline was a wide undifferentiated trough during all the Cambrian and into the earliest Ordovician. In the Lower Ordovician (close of Highgate) came a time of slight orogeny

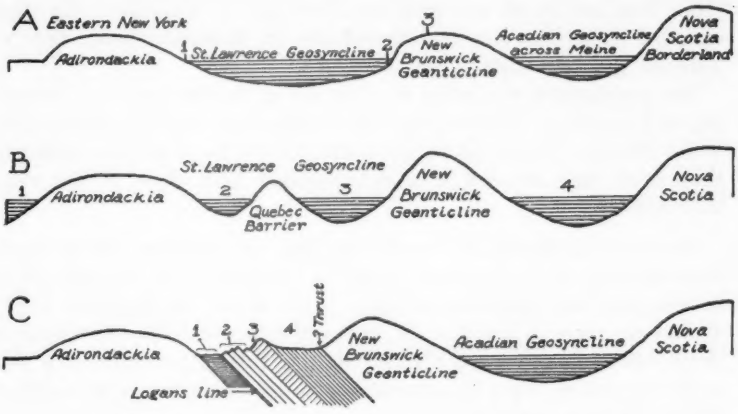


FIGURE 4.—*Theoretic sections across northern New England*

To show general geologic structure at different times (not to scale). (A) Cambrian time: (1) Lake Champlain; (2) Vermont-New Hampshire line; (3) New Hampshire-Maine line. (B) Early Ordovician time, after the Quebec orogeny: (1) Interior seas; (2) Chazy basin with Interior faunas; (3) Lévis trough with Atlantic faunas; (4) Acadian geosyncline with Atlantic faunas. (C) After the Taconian orogeny, in Silurian-Devonian times: (1) horizontal strata of Interior sea; (2) Cambro-Ordovician thrust sheets; (3) Green Mountain pre-Cambrian thrust mass; (4) Cambro-Ordovician folded and metamorphosed strata.

(Quebec phase), which is thought to have made a land ridge near the center of the St. Lawrence geosyncline. To the east of the land ridge lay the Levis trough, with Atlantic faunas; and to the west of it was the Chazy basin, fronting on the foreland of the Canadian Shield and Adirondackia, and marked by Interior faunas. These two Ordovician seaways were blotted out by the Taconian orogeny, which thrust the eastern, or Levis, trough far westward over the Chazy basin.

The present re-elevated Green Mountain-Notre Dame axis seems to have originated during the Taconian orogeny near the close of the Ordovician, and to have been produced by folding and marked overthrusting within the St. Lawrence geosyncline. This deformation made dry land west of the axis, but the trough remaining to the east of it was invaded by the seas of Silurian and early Devonian times, and then was blotted out by the Acadian orogeny.

Since early Middle Devonian time, Vermont has probably been entirely emergent, and the New Brunswick geanticline, during the rest of the Paleozoic, was a highland, which shed immense volumes of detritus into the Appalachian geosyncline from Albany southward, especially during Devonian time. This axis was accentuated by the Acadian disturbance of late Devonian time, which was most intense in Greater Acadia (New England States and Maritime Provinces of Canada). To what extent, if any, this orogeny affected the St. Lawrence geosyncline of Vermont and New York is not yet clear.

The Appalachian revolution near the end of the Paleozoic era, according to Longwell,[25] "deformed the Devonian strata in the southern part of the Hudson Valley; but farther north the results of this movement, if they exist, have not yet been differentiated from those of the older deformations."

Orogeny in southern Quebec.—T. H. Clark published in 1934 an excellent summary of several years' work on the geology of that portion of Quebec that lies immediately north of Vermont. As this area has a similar structure to that part of Vermont with which the present paper deals, and as its geology is representative of that of the entire area west of the core of the Green Mountains, it is thought advisable to present here briefly its main points.

The western front of the Appalachian thrust sheets forms the eastern shore of Missisquoi Bay, striking about N 20° E, and makes "a natural division of the first order. West of this line is the foreland; eastward is a series of thrust slices extending for twenty miles more or less, giving way to the massif of the Sutton Mountains schists, which in turn is followed in the Memphremagog quadrangle by Lower and Middle Paleozoic slates, lavas, and limestones, more or less intensely folded. Thus, a simple four-fold division of the area is easily attained."[26]

The Cambrian and Ordovician formations of the foreland "are for the most part horizontal, but toward the east they have suffered from the pressure of the Appalachian slices moving westward over the thrust planes and are more and more folded, and even crumpled, close to the fault. At the thrust itself the slates are contorted out of all resemblance to the western orderly sequence."[27]

To the east of the slightly moved foreland, the first thrust sheet of the Appalachian Mountains, along the Canadian boundary, is the Philips-

[25] C. R. Longwell, *et al.: Eastern New York and western New England*, 16th Intern. Geol. Cong., Guide Book 1 (1933) p. 5-6.

[26] T. H. Clark: *Structure and stratigraphy of southern Quebec*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 4-5.

[27] *Op. cit.*, p. 5-6.

TABLE 1.—*Formations of the Central Sequence, Northwestern Vermont*

	Taconian orogeny of late Ordovician time.		
	Long break.		
	Grandge slate, at least hundreds of feet thick.		
	Quebec orogeny. Best seen on Grandge farm.		
	Corliss breccia, 30 feet. Known only at type locality.		
Lower Ordovician 1000×feet	Southwest of Milton	Georgia township	Vicinity of Highgate
	?	?	Thrust contact, or Corliss conglomerate, or Grandge slate Highgate slate, 220+feet Highgate limestone, 80 feet Highgate thrust breccia, 25 feet
Upper Cambrian At least 1000 feet		Georgia formation, at least 600 feet	Upper Gorge limestone, 145 feet
		Rockledge breccia, 0–40 feet	Lower Gorge dolomite, 75 feet
		Erosional unconformity	Break and thrust
	Hungerford slate, 600+feet	Hungerford slate, 700 feet	"Hungerford" slate, 200 feet
		Mill River breccia, 20 feet	
	Milton dolomite, 700 feet	Erosional unconformity	"Milton" dolomite, 300 feet
	Break. Some of earliest Upper Cambrian absent.		
	(Lower Cambrian)	(Middle Cambrian. St. Albans)	(Lower Cambrian. Parker)
Mid. Cambrian 270 feet	Late Middle Cambrian. St. Albans slate, 60–270 feet Rugg Brook dolomite breccia, 0–20 feet Break. Most of Middle Cambrian unrepresented		
Lower Cambrian 415–1965 feet	Parker slate, 115–670 feet Short break, at least locally Mallett dolomite and quartzite, 100–630 feet Probable local breaks Winooski dolomite and marble, 200–385 feet Monkton quartzite, 0–300+feet		
	Thrust contact		
	Ordovician basement, Western Sequence, over which the above is thrust		

burg slice, with equivalents of the Lower and the Middle Ordovician (about 5000 feet thick); next eastward is the Rosenberg slice, which has Lower and Middle Cambrian and Lower and Middle Ordovician formations; and then the Oak Hill slice, with Lower Cambrian strata and pre-Cambrian schist. Farthest east is the Sutton Mountains anticlinal belt of low-grade metamorphic rocks.

Toward the close of the Ordovician, Clark says, all the formations were folded and thrust northwest.

"The results of this, the Taconic, orogeny occur over a stretch of country at least sixty miles wide in southern Quebec. [Finally,] there followed a second period of disturbance, the Acadian orogeny. . . . During or immediately succeeding the Acadian orogeny came, it is generally believed, the intrusion of the Stanstead granite."[28]

LOWER CAMBRIAN

FORMATIONS AND FAUNAS

General Statement.—The Lower Cambrian of northwestern Vermont is, in the main, an unbroken sequence, beginning with the Monkton quartzite and continuing upward through the Winooski, Mallett, and Parker formations. The breaks at the base of the Mallett and the Parker are only of short duration. The first two of these divisions were united in the old reports under the name of "Red Sandrock".

Monkton Quartzite.—The red Monkton quartzite[29] is "one of the best key rocks of the region." It is well exposed in the city of Burlington, and the type region is in the township of Monkton, 20 miles to the south. Six miles northwest of Middlebury, it forms the prominent elevation known as Snake Mountain, a remnant of the Central Sequence, which lies on Middle Ordovician shales of the Western Sequence. North of Burlington, the Monkton continues, with the typical expression, into the township of Colchester, but it is generally absent north of Georgia Center, either on account of overthrusting or because of a change in facies.

In its typical development, the Monkton varies in color from reddish-brown through brick-red and purple to light shades of red, pink, buff, and white. The formation is composed chiefly of sandstone, but includes some red or purple shales and, toward the top, grades into the gray or pink dolomites of the Winooski. The only fossils known here are fragments of the trilobite, *"Ptychoparia adamsi,"* which, however, is said to range through the entire Lower Cambrian.

The Monkton formation is essentially one of fine sand, with little mud and dolomite, deposited in a shallow sea with a wide littoral zone; the sandy-beach nature explains the scarcity of fossils. Its thickness, and the formation on which it originally rested, are unknown, as the beds are cut off basally by a major thrust fault, but at least 300 feet has been measured by Keith in Colchester township.

Winooski Dolomite and Marble.—This was probably the first formation in Vermont to receive a name based on a locality; the name appeared

[28] *Op. cit.*, p. 18-19.

[29] Arthur Keith: *Cambrian succession of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 106-108.

CHAMPLAIN THRUST NEAR BURLINGTON, VERMONT

Lone Rock Point. The crumpled Middle Ordovician shale has been eroded back, leaving the resistant Lower Cambrian dolomite (Winooski) as an overhanging shelf; note marked friction channeling at base of thrust plane. Photograph by C. D. Walcott.

FIGURE 1. WINOOSKI MARBLE
Weathered face in Barney quarry, near Swanton (Fig. 1, loc. VI). Note wavy and knotted nature of bedding, which causes mottled red and white blotches when marble is sawed into slabs and polished. Photograph by C. O. Dunbar.

FIGURE 2. HUNGERFORD SLATE
Cleavage crossing bedding, which dips east. West of St. Albans Hill (Fig. 1, traverse XXV). Photograph by C. R. Longwell.

in the literature at least as early as 1861, and its priority in this respect is due to the fact that certain of the mottled dolomites, known in the trade as the Winooski or Swanton mosaic or mottled marbles, have been quarried in the Winooski gorge near Burlington and near Swanton, for about a century. Another trade name for the marble is Wakefield, derived from the quarries on Malletts Bay. The type locality for the geological formation is along the Winooski River and at Winooski Falls, north of Burlington.

It consists of pink, red, mottled, and gray, hard dolomites, more or less siliceous, and showing interbeds of quartzite. The chief character of the dolomite is the wavy, thin, red, impure layers, seen best on weathered surfaces (Pl. 4, fig. 1). The mottling of the marble is due to the shallowness of the sea in which the beds were laid down, and to the rippled nature of the deposits. The thickness varies between 200 and 385 feet (12 miles north of Burlington); the greater thicknesses given by authors (810 to 970 feet) include Mallett and Monkton. South of Burlington and north of Swanton the Winooski grades downward into the Monkton.

The marble beds carry rare *Salterella pulchella* (? primitive cephalopods), and Edson [30] reports having seen in addition *"Ptychoparia adamsi"*; Resser [31] records *Bonniella desiderata.* A little west and south of Parkers Cobble, in the cliff facing Lake Champlain, and in what appears to be the Winooski formation, Walcott discovered 50 feet of buff calcareous sandstone, which has, toward the top, *Obolus* near *O. crassus, Botsfordia caelata, Hyolithes americanus, H. communis, Hyolithellus micans,* fragments of *Olenellus,* and a head of *Ptychoparia* near *P. trilineata* (Walcott's localities 319j and 319k).[32] *Salterella* is found at horizons 85 feet, 113 feet, and 198 feet above the buff sandstone, in decomposed limestones of the Winooski.

Contact Between Winooski and Mallett.—It is generally assumed that deposition was continuous throughout Winooski and Mallett times, and this may be the correct view. The writer has walked with Keith over the contacts at several places, and the deposition appears so regular that nowhere was a marked break seen. On the other hand, the seas of those times were shallow, and intraformational conglomerates are not rare, hence, local breaks and such conglomerates are to be expected. Howell [33] describes such a break.

[30] G. E. Edson: *Geology of the town of Swanton, Vt.,* Report of the State Geologist . . . of Vermont, 1907-1908 (1908) p. 218.

[31] C. E. Resser: *Third contribution to nomenclature of Cambrian trilobites,* Smithsonian Misc. Coll., vol. 95, no. 22 (1937) p. 5.

[32] C. D. Walcott: *Cambrian Brachiopoda,* U. S. Geol. Surv., Mon. 51 (1912) p. 251.

[33] B. F. Howell: *Contact of the Cambrian Winooski and Mallett formations of northwestern Vermont,* Geol. Soc. Am., Pr., 1933 (1934) p. 333.

Mallett Dolomite.—The type locality for this formation [84] is on Malletts Bay of Lake Champlain, 5 miles north of Burlington, where it is 800 feet thick; elsewhere the thickness is limited by overthrusting to from 100 to 630 feet, but generally it is less than 200 feet. The formation continues north across the Canadian boundary for a few miles. It consists chiefly of massive, light- and dark-gray sandy dolomites, but north of St. Albans as much as one-sixth of it is made up of fine-grained, more or

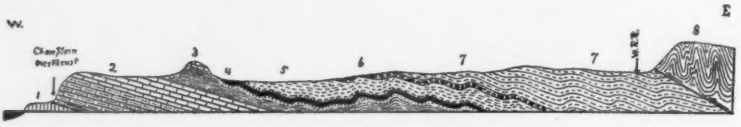


FIGURE 5.—*The celebrated Georgia section*

From Lake Champlain east to Vermont Central Railroad (Fig. 1, loc. IV). (2-7) Central Sequence, overthrust on (1) of Western Sequence. (8) Eastern Sequence, overthrust on (7) Georgia formation of Keith. Lower Cambrian: (2) Winooski-Mallett dolomite; (3) Parker slate with dolomite capping. No Middle Cambrian present. Upper Cambrian: (4) basal Hungerford slate; (5) main mass of Hungerford slate; (6) Rockledge breccia at base of (7) Georgia formation of Keith.

less cross-bedded quartzites. The sand in these beds consists of glassy, rounded quartz grains which stand out in relief on weathered surfaces. This same kind of sand, however, is seen in all subsequent dolomites and limestones, and in the paste of the limestone conglomerates as well. It appears to be wind-blown sand, originally derived from the pre-Cambrian of the foreland, subsequently weathered out of Lower Cambrian deposits, and re-worked into younger Cambrian and Ordovician formations. Another feature of the Mallett sandstone is rippling, with channeling and the introduction of well-rounded or angular intraformational pieces of dolomite, some of which are 18 inches across. This is further evidence of shallow seas and tidal-current action. Shale beds are rare in the Mallett, and a sandy intraformational dolomite breccia, 30 feet thick, is exposed locally near the base. The contact with the overlying Parker shale is generally sharp, indicating a break in sedimentation (Fig. 8).

Fossils are exceedingly rare in the Mallett dolomite, and most of those recovered are from decomposed calcareous sandstones. East of Highgate Springs, Walcott reports *Palaeophycus incipiens, Paterina swantonensis, Nisusia festinata, Scenella varians, Ptychoparella walcotti, P. swantonensis, P. billingsi, P. adamsi, Antagmus typicalis, Hyolithellus micans,* and *Olenellus thompsoni.*

West of Parkers Cobble, Walcott found *Paterina swantonensis, Hyolithellus micans, Salterella pulchella, "Ptychoparia adamsi",* and *Olenellus thompsoni.*

[84] Arthur Keith: *op. cit.*, p. 110.

Break Between Mallett and Parker.—At least one known place records a distinct erosion interval between the deposition of the Mallett dolomite and that of the Parker slate (Fig. 8). This locality is described on page 1034. In several places the basal beds of the Parker slate have dolomite inclusions, but actual basal conglomerates were not seen. On the other hand, the fossils known from the two formations are not different, indicating that the erosion interval was not a long one.

Parker Slate.—Marcou, in 1861, called this formation Georgia slate, and Walcott used the same name in 1886. Keith in 1923 [35] re-named it Colchester shale, but nine years later [36] again changed the name, because of the poor exposure of the formation in Colchester and the absence of fossils there. Keith took the present term from Parkers Cobble on the old Noah Parker farm, the type locality for most of the fossils described by Walcott; the farm is 2 miles N 60° W of Georgia Center (Fig. 1, loc. XVI). The typical section is exposed around the Cobble on the Parker farm and for some miles to the south and north along the strike of the formation. In this region, the thickness is approximately 300 feet. Elsewhere, the thickness ranges between 115 and 670 feet, but is generally less than 300 feet. Much farther north, near Swanton, the thickness is between 340 and 675 feet. These variations are due to the erosion of the top of the formation.

The Parker strata are predominantly dark-blue micaceous shale, with some flaggy sandstone in the lower part. There is a sharp change, generally without transition, from the Mallett dolomite to the Parker slate, and one place shows a distinct erosion interval. The upper Parker also has local zones of dolomite, the thickest one, about 30 feet, capping Parkers Cobble. (See also p. 1013 and 1035.) In the township of Swanton, these recurring local dolomites are more common than elsewhere. To the south of Burlington, sandy dolomites and calcareous sandstones become more numerous.

In addition, the Parker slate has local reef limestones (bioherms), "consisting of massive blue marbleized limestone, sharply separated from the slate. They are best shown two miles southeast of Swanton, where two of them are surrounded by gray slate [Fig. 1, loc. XVIII]. The lenses measure 100 by 60 feet and 105 by 75 feet. . . . This limestone very closely resembles some of the bowlders in the later conglomerates." [37]

The Parker slate is the only Lower Cambrian formation that is in

[35] *Op. cit.*, p. 110-112.

[36] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 371-372.

[37] Arthur Keith: *Cambrian succession of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 111.

places richly fossiliferous, but good material is hard to get. On the Parker farm (Fig. 1, loc. XVI), there is little cleavage, and most of the fossils collected for Walcott were obtained there by quarrying in two different places. From one place or another, the following species have been described: *Palaeophycus incipiens, P. congregatus, Dactyloidites asteroides, Rustella edsoni, Paterina labradorica, Micromitra (Paterina) bella, Kutorgina cingulata, Nisusia orientalis, N. festinata, Microdiscus parkeri, Mesonacis vermontana, Olenellus thompsoni* and var. *crassimarginatus, Paedeumias transitans, Olenoides marcoui, Bathynotus holopyga, Ptychoparella vermontensis, P. georgiensis, Periomma typicalis, Bonnia swantonensis, Protypus hitchcocki,* and *Protocaris marshi.*

Long Erosion Interval.—During Lower Cambrian time, between 1500 and 1800 feet of sediments were laid down, indicating the amount of subsidence in the St. Lawrence geosyncline. Then came a long-continued emergence of the region, without crustal disturbance, and, during this erosion interval, much of the Parker slate was removed.

Significance of the Dissimilar Cambrian Faunas in Northwestern Vermont and in the Taconic Mountains.—The region from Snake Mountain south to Brandon, in the area of the Taconic Mountains, does not furnish Lower Cambrian fossils. To the south of Brandon, beginning with West Castleton, the fossils and lithology are clearly those of eastern New York, and the region is characterized by the trilobite *Eodiscus speciosus.*

To the north of Snake Mountain, the lithology is of fairly uniform character north to the Canadian boundary, and the Lower Cambrian fauna of this area, although not wholly unlike that of the Taconic Mountains, yet has its own impress and is known to extend northeast to southeastern Labrador and northeastern Newfoundland. Probably the most characteristic fossils of this area are *Olenoides marcoui, Ptychoparella adamsi, Mesonacis vermontana, Bathynotus holopyga, Paterina labradorica,* and *Kutorgina cingulata.* That the seas of these two areas were in communication, however, is proved by the twelve species which they have in common. Accordingly, it is probable that this difference in faunas reflects a difference in sedimentary facies, for the sea of the Taconic area laid down mainly muds, whereas the northern one deposited mostly dolomites, with a minimum of sands.

To bring out more clearly the marked faunal differences between northwestern Vermont and the northern Taconic Mountains, Walcott's list [88] of 45 species from the latter region is here presented, an asterisk marking the twelve species that occur also in northwestern Vermont:

[88] C. D. Walcott: *Cambrian Brachiopoda,* U. S. Geol. Surv., Mon. 51 (1912). List of localities, p. 161-291.

Archaeocyathus dwighti
A. rarus
A. rensselaericus
**Paterina swantonensis*
Micromitra pannula
Obolella chromatica
**O. crassa*
O. minor
Obolus prindlei
**Lingulella granvillensis*
L. schucherti
**Botsfordia caelata*
Bicia gemma
B. whiteavesi
Yorkia washingtonensis
Acrotreta emmonsi
**A. sagittalis taconica*
Acrothele nitida
A. pretiosa
**Nisusia festinata*
Billingsella salemensis
Scenella retusa
Stenotheca elongata
**S. rugosa*
Platyceras primaevum
**Hyolithes americanus*
**H. communis*
H. impar
**Hyolithellus micans*
Fordilla troyensis
Leperditia dermatoides
Agnostus desideratus
Eodiscus speciosus
Microdiscus connexus
M. lobatus
M. schucherti
Solenopleura nana
S. tumida
Protypus clavatus
**P. hitchcocki*
Zacanthoides eatoni
Olenoides fordi
Ptychoparia fitchi
**Olenellus thompsoni*
Elliptocephala asaphoides

In order to understand what these facies differences mean, the broader structural conditions must also be considered. These show the astonishing fact that the formations of the Taconic Mountains, now in southwestern Vermont and eastern New York, seemingly must have originated far to the east, even beyond the present Green Mountains, and were, it is thought, the near-shore eastern facies of the St. Lawrence geosyncline.

The rocks of the Taconic sequence, Keith states, "form a great mass that was thrust westward from the region far to the east over the rocks of the eastern sequence and later isolated by erosion." [39]

According to Prindle and Knopf,[40] the Taconic Mountains, a synclinorium, extend from Orwell and Sudbury, Vermont, southward for 200 miles into New York to the east of the Hudson River.

"It is now reasonably well established that the main structure of the Taconic Mountains is an overthrust mass that has been packed into tight folds in a general synclinal structure. . . . The eastern flank of the Green Mountains is the only place where such a series of argillaceous rocks as the Taconic sequence could have been deposited contemporaneously with the early Paleozoic sandstone and carbonate facies now exposed on the west flank of the Green Mountains. Whether this argillaceous sequence was transported from its original position westward along a low-angle thrust or by a great series of recumbent folds rolling one above the other and intensified by low-angle thrusts must remain to a large extent a matter of speculation." [41]

[39] Arthur Keith, *in* C. R. Longwell, *et al.: Eastern New York and western New England*, 16th Intern. Geol. Cong., Guide Book 1 (1933) p. 61. Washington.

[40] L. M. Prindle and E. B. Knopf: *Geology of the Taconic quadrangle*, Am. Jour. Sci., 5th ser., vol. 24 (1932) p. 260-261, 297, 301.

[41] *Op. cit.*, p. 297.

Prindle and Knopf illustrate in their figure 4 the probable general structure of the Green and Taconic mountains as revealed in the Bennington, Greylock, Hoosick, and Berlin quadrangles.

The interpretation suggested calls for a westward transportation of at least 40 miles and possibly as much as 70 miles. Accordingly, the fossils of the Taconic Mountains originally lived along the eastern shore of the St. Lawrence geosyncline, whereas those of northwestern Vermont probably lived toward the center of this same seaway or much nearer the western shore. The differences in these faunas are, then, the result of geographic and facies conditions.

DETAILED DESCRIPTION OF LOCALITIES

Northeast of Highgate Springs.—This is the farthest north of the localities to be described. A section of 1125 feet was measured by the writer in 1922, east of Rock River and about 2 miles northeast of Highgate Springs (Fig. 1, loc. XIX). The ascending series begins at the west, near the Swanton-St. Armand road, where the sequence is overthrust on the Philipsburg block,[42] and goes east to Saxe Brook. The section is as follows:

MONKTON FORMATION (125-150 feet).

Red, fine-grained, massive, and evenly bedded quartzite, with thin partings of red shale showing fucoidal markings and fragments of trilobites. The writer regards these strata as of Monkton time, but others add them to the Winooski.

WINOOSKI FORMATION (346 feet).

Bed		Thickness (Feet)
1.	Light-gray to pink, thick-bedded dolomite	60
2.	White, massive, coarse to conglomeratic quartzite	10
3.	Gray, fine-grained, sandy dolomite	8
4.	Gray, coarse-grained quartzite	30
5.	Dark-red, fine-grained quartzite	15
6.	White, fine-grained, dolomitic quartzite, poorly exposed	25
7.	Light-gray, heavy-bedded, cliff-making dolomite	50
8.	Gray sandy dolomite, poorly exposed at farm fence	50
9.	Gray dolomite, with some rounded sand grains	8
10.	Gray sandy dolomite below, passing upward into coarse sandstone	8
11.	Red sandy dolomite	3
12.	Gray dolomite in thick beds, with some galena	20
13.	Red, dolomitic, coarse sandstone	6
14.	Pink, dolomitic, coarse sandstone	5
15.	Covered zone	10
16.	Coarse sandstone	2
17.	Gray dolomite	5
18.	Gray, even-bedded sandstone, with *Ptychoparella adamsi*	5
19.	Pink, fine-grained sandstone	4
20.	Pink to red, fine-grained dolomite	10
21.	Red, thin-bedded and laminated, fine-grained sandstone, with an intraformational conglomerate of the same bed	12

[42] H. W. McGerrigle: *Philipsburg series of southern Quebec*, Geol. Soc. Am., Bull., vol. 42 (1931) p. 347-348.

McGerrigle recognized no Monkton or Winooski north of the Missisquoi River, although, east of Highgate Springs and for 2 miles north, he did see reddish beds like the Monkton. These beds have *"Ptychoparia adamsi"* and "are interstratified with brown-weathering, grayish dolomites through a thickness of about 200 feet, after or above which the red beds disappear but the dolomites continue and become interbedded with heavy layers of grayish quartzite. Thus, the 'red beds' seem to be interstratified with, and to be conformably overlain by dolomites of the Mallett type without the intervention of the Winooski."[43]

McGerrigle has mapped these 200 feet of strata with the Mallett, but the writer refers them to the Winooski; 5½ miles south, the Winooski marble is present in the Barney quarry.

The section above the Winooski is:

MALLETT FORMATION (631 feet—an unusual thickness, due either to local thrusts or to the inclusion of equivalents of the Winooski formation).

Bed	Thickness (Feet)
1. Pink, fine-grained dolomite, with the basal 5 feet an intraformational breccia of angular blocks as much as 2 feet long	25
2. Light-pink, sandy dolomite. In all these sandy dolomites the grains are well-rounded, wind-blown sand	4
3. Gray, fine-grained dolomites (10 feet); above, a covered zone (10 feet)	20
4. Yellowish-gray, massive dolomite	25
5. Coarse sandy dolomite	25
6. Gray, fine-grained dolomite, with a few zones of sandy dolomite	130
7. Yellowish-gray, coarse-grained dolomite	10
8. Yellowish, decidedly cross-bedded quartz conglomerate	5
9. Fine-grained, banded quartzite	4
10. Decidedly cross-bedded quartz conglomerate. Channeled into the bed below to a depth of 12 inches	5
11. Gray, fine-grained, decidedly sandy dolomite	12
12. Gray, fine-grained, cross-bedded quartzite	8
13. Yellowish sandstone, with dolomitic cement	28
14. Decidedly cross-bedded and channeled coarse quartzite	5
15. Covered interval	20
16. Coarse sandy dolomite, with interbedded quartzite	12
17. Covered area at road and fence	20
18. Gray, coarse sandy dolomite (10 feet), and gray, fine-grained dolomite (10 feet)	20
19. Bedded, rusty sandstone, with fragments of fossils	3
20. Gray, fine-grained dolomite	5
21. Well-bedded, fine-grained sandstone	5
22. Cross-bedded quartz conglomerate (2 feet), gray, fine-grained, regularly bedded sandstone (3 feet)	5
23. Covered zone	5
24. Well-bedded sandy dolomite	20
25. White, massive, dense quartzite	10
26. Coarse to conglomeratic, much cross-bedded sandstone, with irregular layers of sandy dolomite	5

[43] H. W. McGerrigle: *Three geological series in northwestern Vermont*, Vt. State Geol., 17th Bienn. Rept., 1929-1930 (1931) p. 183.

Bed		Thickness (Feet)
27.	White, massive, dense quartzite. The base is channeled into the bed below and has many rounded pebbles, 2 to 6 inches long, from that bed; also an angular block, 18 inches long	8
28.	Coarse-grained, decidedly cross-bedded sandstone	5
29.	Yellowish-gray sandy dolomite	30
30.	White quartzite	2
31.	Massive and evenly bedded dolomite (35 feet); gray, fine-grained dolomite (30 feet)	65
32.	Sandy dolomite, massive and cross-bedded	40
33.	White, fine-grained quartzite (5 feet), sandy dolomite (5 feet), white cross-bedded quartzite (5 feet)	15
34.	Sandy dolomite	25
35.	Quartz conglomerate, with pebbles up to ¼ inch	5

Farther east is the flat of Saxe Brook, some hundreds of feet across, with the stream flowing over Champlain clay, which probably conceals Parker slate eastward to an escarpment, which may be of overthrust Mallett. As no fossils are known from this dolomite, the question arises whether it is Mallett or the "Milton" dolomite seen farther southeast (p. 1031).

East of Highgate Springs.—The oldest beds in this section (Fig. 1, traverse XX) are on the east side of Carman Brook and east of the road from Swanton, Vermont, to St. Armand, Quebec; the section thence ascends eastward into younger strata. The following description, however, which is after Walcott,[44] begins in the east and goes westward into older strata:

MALLETT FORMATION (200 feet)

Bed		Thickness (Feet)
"1.	Compact gray silicious limestone in massive layers	20
"2.	Gray calciferous limestone, compact, hard, evenly bedded, breaking up into shaly layers in places. Some of the layers decompose on exposure into a reddish sandy rock	180"

WINOOSKI FORMATION (970 feet). Contact with Mallett uncertain.

Bed		Thickness (Feet)
"3.	Compact purplish, pinkish, or greenish colored silicious limestone in massive layers	30
"4.	Thick layers of buff and pinkish-colored silicious limestone with fragments of *Ptychoparia adamsi*	15
"5.	Heavy-bedded, reddish-purple, fine-grained magnesian limestone, breaking into angular fragments (*Ptychoparia adamsi* in abundance). Strike N. 30° E., dip 25° to 28° E.	50
"6.	Shaly and massive layers of gray and purplish colored sandstones containing fragments of *Ptychoparia adamsi*, occurring at various horizons	175
"7.	Reddish-colored arenaceous limestone, with irregularly bedded massive layers of gray silicious limestone and a few more-evenly-bedded layers carrying fossils." With beds of gray and buff sandstone. *Paterina labradorica* (= *swantonensis*), *Nisusia festinata*, *Scenella varians*, *Olenellus thompsoni*	700"

[44] C. D. Walcott: *Second contribution to the studies on the Cambrian faunas of North America*, U. S. Geol. Surv., Bull. 30 (1886) p. 18-19.

Two miles North 15° West of Highgate Center.—Howell, in 1934, took the writer to this place (Fig. 1, black square of traverse V), on the eastern side of Saxe Valley and east of the north-south road. On the rising ground toward the "Mallett" dolomite, the Parker slate shows beds of dark-blue sandy limestones, which weather into reddish, porous, soft rock having fragments of *Olenoides marcoui, O. quadriceps, Bonnia, Ptychoparella,* and *Coleoloides?*. In the interbedded slate was found an entire specimen of *Ptychoparella adamsi?*.

FIGURE 6.—*Sequence from Russell quarry west*

Distance, about 2 miles (Fig. 1, traverse V). (1) Philipsburg series overthrust; (2-3) Winooski and Mallett dolomites; (4) almost horizontal Parker slate across a wide valley; (5) fossiliferous Parker slate; (6) "Milton" Upper Cambrian dolomite, formerly called Mallett; (7) "Hungerford" slate; (8) Gorge dolomite.

The Saxe Valley, here about half a mile wide, is believed to be floored entirely with horizontal Parker slate, because to the west is rising land with undoubted Mallett and Winooski formations. If the dip is constant at 10° E., the Parker may be 350 feet or more in thickness (Fig. 6).

This locality is important because the region has been mapped by McGerrigle, who says:

"As this road is followed northward the Colchester [= Parker] outcrops sparsely and intermittently on the eastern side for about two miles. In this distance the Milton dolomite [= what is referred to by the present writer as "Mallett"] lies across the line of strike of the Colchester [= Parker], apparently having attained this position through thrusting."[45]

If this "Milton-Mallett" has been overthrust upon the Parker, which is surely present in Saxe Valley, then the dolomite should be correlated with the Mallett. A question then arises about the overlying slate, the top of which is seen in the Russell quarry, where its fossils, once thought to be Lower Cambrian, now are seen to indicate an age not older than Upper Cambrian. Relying upon the fossils found above and beneath the dolomite, the writer believes that the "Milton-Mallett" dolomite is not in thrust relation upon the Parker beneath, but that, subsequent to its erosion, the older slate was disconformably overlapped by the "Milton" dolomite, now referred by Howell and the writer to the Upper Cambrian (see p. 1057).

Southeast of Swanton village.—The following section (Fig. 1, traverse VI) is the classical one of Logan and Billings, and is described by Logan[46] as being "over a mile east of Swanton Falls" of the Missisquoi River, but at the place indicated the beds are either Lower Ordovician of the Western Sequence, or Champlain clays. The nearest place for the Lower Cambrian is about 1.5 miles southeast of the village of Swanton, at the old "Barney quarry" in Swanton mottled red and white marble, on the south side of the Missisquoi River. East of the quarry, the descending sequence as given by Logan is as follows (Fig. 7):

[45] H. W. McGerrigle: *op. cit.*, p. 189.
[46] W. E. Logan: *Geology of Canada* (1863) p. 281-282.

PARKER SLATE (190+ feet; east of the Barney quarry, there is about 340 feet of slate).

Bed	Thickness (Feet)
"5. Measures not well exposed" of bluish slate with zones of limestone, dolomite, and micaceous flagstones	60
"4. Dark grey and bluish-black slate, partially magnesian, with thin bands of sandy dolomite, weathering yellowish. The slate is fossiliferous, but, having a cleavage independent of the bedding, it is somewhat difficult to obtain specimens." Toward the top occur *Kutorgina cingulata, Nisusia festinata, Swantonia antiquata, Antagmus teucer, Olenellus thompsoni,* and *Mesonacis vermontana*	130

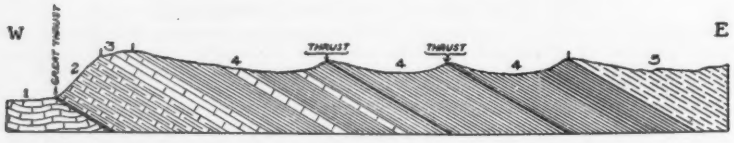


FIGURE 7.—*Sequence from Swanton marble quarry to Donaldson farm*

Distance, 1.25 miles (Fig. 1, traverse VI). (1) Western Sequence; (2-5) overthrust Central Sequence, with one major and two minor thrusts. Lower Cambrian: (2) Winooski dolomite; (3) Mallett dolomite; (4) Parker slate; (5) Parker slate of Donaldson farm. For continuation of this sequence, see Figure 12.

Divisions 4 and 5 of Logan are now little exposed east of the Barney quarry, and what can be seen is as follows: Along the short-cut telegraph line through the woods and swamps, two hard ridge-making layers of the Parker formation are exposed at the extreme east. To the east of these hard beds, there is a wide expanse of Lower Cambrian formations, extending eastward about a mile to the Donaldson farm, where the Parker slate is well exposed and has long yielded the characteristic fossils of the formation (*see* under Donaldson farm, the next locality to be described). Farther north, this wide expanse of Parker slate is known to be either in undulations with low dips, or thrustfaulted. Therefore, without traversing and studying the structure of this area in detail, one cannot say how much more, if any, must be added to the ascertained thickness of about 340 feet for the Parker.

Returning west along the telegraph line, one soon encounters a ridge of pink limestone interbedded with mudstones, entirely unlike anything that has been seen elsewhere in the Lower Cambrian. The beds are 15 feet thick and have a dip of 9 degrees to the east. Trilobite fragments are common here, but the rolling of the limestone into nodules has so broken and mixed them that the only determinable fossil collected is *Olenoides marcoui?*.

The next exposure, about 750 feet west, shows a yellowish gray, coarsely crystalline dolomite, about 20 feet thick, with no fossils. Westward, the beds are again covered for 1100 feet, and then low cliffs expose about 35 feet of Parker slate. Still farther west for 500 feet is another covered area, which appears to have both Parker slate and Mallett dolomite. Eliminating this last covered zone, the thickness of the Parker slate is about 340 feet.

The descending section of Logan continues west as follows:

MALLETT DOLOMITE (190 feet).

Bed		Thickness (Feet)
	"3. Buff sandy dolomite, weathering brown (40 feet)." To the east of this well-exposed zone is a covered area, 500 feet across, which shows more or less of the same Mallett sandy dolomite, giving a thickness for all of	80
	"2. Grey argillaceous limestone, partially magnesian, weathering brownish and holding a great abundance of tangled specimens of *Palaeophycus incipiens.*" This is the transition zone to the older Winooski	110
	The upper half of this zone is well exposed in a cliff, and about 15 feet of it is slabby on weathering, dark-blue, and in appearance suggests the Parker slate, for which it was mistaken by Edson.[47] It is a hard massive magnesian series, replete with the fucoid, but has no other fossils.	

WINOOSKI DOLOMITE (370 feet), extending east of the quarries and beneath them to near the level of the river.

Bed

"1. Gray, blue, and red dolomites, with sandy layers and red mud partings." Some of the strata are mottled rose-red and white, and a few are brick-red or Indian red. Other beds have considerable siliceous sand, and all weather to a reddish or yellowish brown. Certain red beds contain *Ptychoparella adamsi* and *P. vulcanus.* It is in the lower third that the Swanton marble is found.

Regarding this section, Logan[48] says that the Parker slates have been followed southward to similar beds at Parkers Cobble. Northward, they can be traced to Saxe's mill on Saxe Brook. Billings, with the help of local collectors, gathered the fossils "in the black slates" in the region of the Swanton marble quarry, and in 1861 described *Palaeophycus incipiens, Kutorgina cingulata, Nisusia festinata, Swantonia antiquata, Olenellus thompsoni,* and *Mesonacis vermontana.*

East of the Swanton marble quarry, according to Edson,[49] are "lenticular masses of limestone which is very rich in fossils". At this place he collected *Kutorgina cingulata, Nisusia festinata, Paterina swantonensis, Rustella edsoni, Swantonia antiquata, "Ptychoparia adamsi," Olenellus thompsoni,* and *Mesonacis vermontana.* The writer has not been able to find this lenticular limestone, and the much higher one with *Olenoides marcoui,* which he did find, does not have the fossils mentioned by Edson.

About a mile east of Highgate Springs, Billings found *Labradoria misera;* and in flaggy sandstone beside the road leading from Moores Corners, in St. Armand, to Saxe's mill in Highgate township, about a mile south of the Canadian line, he collected *Perimetopus arenosus* and *Ptychoparella ? vulcanus.*

These remarkable field and fossil discoveries, so clearly described in 1861 and 1863 by Billings and Logan, should have quieted the Taconic controversy, especially as Barrande had said that the fossils were of his Primordial system. However, the controversy continued long afterward, because Hall could not divest him-

[47] G. E. Edson: *The geology of St. Albans and vicinity,* Vt. State Geol., 5th Biennial Rept., 1905-1906 (1906) p. 153.
[48] W. E. Logan: *op. cit.,* p. 282.
[49] G. E. Edson: *op. cit.,* p. 153.

self of his wrong conclusions as to the age of the fossils which he had earlier described.

Donaldson farm, east of Swanton.—The Donaldson farm is about 6.5 miles north of St. Albans and thence less than a mile to the west on a farm road (Fig. 1, loc. XVIII). The section of the Cambrian to the east is described on pages 1066-1067, and that to the west on pages 1032-1033. The Parker slate is exposed on the eastward-descending dip slope, and has a bed rich in Lower Cambrian fossils. The writer was informed in 1922 that about 1881 or 1882 "an old man, a young man,

FIGURE 8.—*Erosion contact between Parker slate and Mallett dolomite*
Lower Cambrian. North of Swanton Junction (Fig. 1, loc. VII). This exposure is included in Figure 9.

and a boy" collected fossils on this farm. These people were Mr. Rust (a collector), Mr. Walcott, and probably the latter's son, Charles. The place appears to be the old G. M. Hall locality recorded by Logan,[50] and is Walcott's locality 25a.[51] From it the latter records *Paterina swantonensis, Lingulella granvillensis, Kutorgina cingulata, Nisusia festinata, N. orientalis, Salterella pulchella, Olenellus thompsoni, "Ptychoparia adamsi,"* and *Bonnia*. The writer did not collect all these species, but found in addition *Hyolithes americanus, Stenotheca rugosa,* and *Scenella* sp.

West of Donaldson's farm, the Parker consists of dark-blue, evenly bedded and banded, sandy and flaggy micaceous slates. These flags can be picked up in places in sheets 3 feet across, and many of them are also to be seen in the stone walls. Some of the slabs show fragments of *Olenellus* and specimens of *Hyolithes*.

On the southeastern slope of the Donaldson farm, and above the Parker slate, lie remnants of the much younger Hungerford basal thrust breccia and a bioherm out of the Hungerford slate, about 105 feet long. There is another blue-gray lentil near the Donaldson home, but this one is in the Parker slate; it is about 65 feet long and projects about 20 feet above the slate.

Another place for fossils in this region is Walcott's locality 87, less than a mile northeast of the Donaldson farm, which has yielded *Paterina swantonensis* and *Lingulella franklinensis* (type locality).[53]

Three miles southeast of Swanton.—Less than a mile north of Swanton Junction (Fig. 1, loc. VII), a significant contact between the Parker slate and the Mallett dolomite is to be seen. The dip of the beds is about 10° E. The Mallett was eroded, before Parker time, into an uneven surface, with depressions 10 feet deep, and over this irregular surface the Parker slate was deposited (Fig. 8). This locality also helps to explain what is seen at several others—namely, that the basal 5 to 10 feet of the Parker slate is more or less conglomeratic, the pebbles being small angular pieces of dolomite, which weather out, leaving iron-rust cavities.

[50] W. E. Logan: *op. cit.*, p. 280.
[51] C. D. Walcott: *Cambrian Brachiopoda*, U. S. Geol. Surv., Mon. 51 (1912) p. 188.
[53] *Op. cit.*, p. 217.

The slates continue eastward from the contact just described, and the road-metal quarry of J. P. Kelly (Fig. 1, loc. XXI) exposes less than 30 feet of Parker slate, dipping 10° E. The total thickness of the Parker here is estimated at 180 feet (Fig. 9). In the lower part of the quarry, which has long been known as a locality for Lower Cambrian fossils, an entire but uncollectable specimen of a small *Mesonacis vermontana* was seen. Near the entrance may be seen square yards of slate covered with *Dactyloidites asteroides*, discovered by Edson in 1906. In the higher beds of the quarry occur *Ptychoparella adamsi, Acrocephalites ? vulcanus*, and *Microdiscus*.

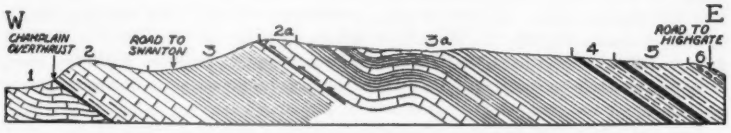


FIGURE 9.—*Sequence north of Swanton Junction station*

Beginning 1 mile north of the station and extending southeast less than 2 miles to Highgate road (Fig. 1, traverse VIII). (1) Ordovician of Western Sequence, over which is thrust the Central Sequence as follows: (2) Lower Cambrian Winooski-Mallett, followed by (3) Parker slate (see contact in Figure 8), which is terminated by dolomite, 25 feet thick. (2a) A minor thrust of Mallett, 75 feet thick, followed by (3a) Parker with dolomite zones. (4) Middle Cambrian, absent in this section but occurring less than a mile south. (5) Upper Cambrian Hungerford slate with basal Mill River breccia; (6) Rockledge breccia.

Above the quarry, one finds from 10 to 15 feet of calcareous slates, weathering rusty red, some of which have an abundance of fragmented *Olenoides marcoui*, rarer fragments of *Olenellus thompsoni, Hyolithes edsoni*, and a new brachiopod nearest to *Kutorgina*. The calcareous slates are overlain without break by about 25 feet of ridge-making, somewhat conglomeratic (intraformational) dolomite, dove-colored when fresh but weathering to salmon color; this appears to be the same dolomite that makes the Cobble on the Parker farm. The beds dip to the east. Beyond a little hollow, another ridge of dolomite exposes 75 feet of strata (Fig. 9). The question now arises, Is the sequence here continuous, or is it broken by an overthrust between the two dolomites? Several minor factors appear to indicate that the second dolomite is overthrust Mallett, and this is also Keith's view.

Above the overthrust Mallett, eastward (Fig. 1, traverse VIII) for about 3600 feet, a series of outcrops exposes Parker slate, in undulations, with many short and thin lenticular beds of dolomites, which weather rusty red, and one bluish white dolomite, 20 feet thick, which becomes increasingly sandy and coarse-grained upward. The average dip of these strata is probably less than 10° E., but in one place a low anticline shows dips up to 35° W. Their estimated thickness is about 675 feet, the maximum for the Parker slate. The only fossil recovered, a large glabella of *Microdiscus*, was found near the base. The Parker slate is overlain by dolomite conglomerate that makes the base of the Upper Cambrian. No Middle Cambrian occurs here (Fig. 9). The eastward continuation of this section is given on page 1065.

From the foregoing account, it is clear that this section of the Parker slate and the one southeast of Swanton (Barney quarry) have more dolomite than is the rule south of St. Albans.

Near Fonda Quarry, south of Swanton Junction.—A short distance north of Fonda quarry, on either side of a cross-road that goes east from the main Swanton highway

to the Rockledge estate, many exposures of the Lower Cambrian are to be seen (Fig. 1, traverse XXII). Beginning at the west, the Winooski is buried, but it is exposed a short distance to the north, on the eastern side of the highway to Swanton. Then eastward on the road to Rockledge, gray lower Mallett and the transition shaly strata to the gray upper Mallett dolomite are exposed. Above the latter is a nearly complete section of the Parker slate, which is more or less crumpled, probably on account of the overriding of another thrust sheet, which repeats more or less of it. To the north of the third house from the west is a quarried cliff of the Parker slate, dipping only a little to the east, over which lies a dolomite zone, 10 to 15 feet thick, considerable parts of which are brecciated and shoved into the adjacent slate. A short distance to the east of this house the gray Mallett is also exposed, dipping 10° E., and evidently making the base of a higher and greater thrust sheet within the Central Sequence. Over the final Mallett outcrop is a long succession of the Parker slate, having four zones of dolomites, which weather rusty red and vary in thickness from 5 to 15 feet.

Across the Swanton road east of the Fonda quarry is the old Bullard farm noted by Logan,[53] where the Parker slate is also exposed. Walcott[54] cites this place as his locality 319r, which yielded him *Rustella edsoni* and *Olenellus thompsoni*.

North of St. Albans Bay.—A short distance to the north of the road leading from St. Albans to St. Albans Bay (Fig. 1, traverse II), a low knoll exposes the marble horizon of the Highgate Springs series (Middle Ordovician) of the Western Sequence (Fig. 2, C, bed 10). In this vicinity, but to the south of the St. Albans road, other small exposures of light-blue limestones of the Highgate Springs series, with indistinct traces of gastropods, are to be seen; the upper part of this same series is present in another hillock to the south of the next east-west road, where it contains a *Phylloporina*, two *Maclurites* and opercula, other gastropods, *Orthoceras*, and other forms. These fossils are poor, but they appear to be of Chazy age.

Over the marbles of the Western Sequence already described lie the overriding Winooski dolomites, to be seen in many exposures on a rising hilly land to the east, but no actual thrust contact between them is exposed. These dolomites dip to the east quite regularly about 9 degrees, and contain *Palaeophyclus* at the base. The Winooski at this place probably does not much exceed a thickness of 200 feet. Accordingly, only the upper part of this formation is present, with all of the lower Winooski and the Monkton faulted out (Fig. 2, C, bed 4).

No transition zone exists here from the Winooski to the Mallett dolomite, such as is present near Swanton, but with the introduction of the Mallett, wind-blown etched sand begins to appear. Near the base of the Mallett, there is a zone of massive, intraformational, sandy, dolomitic breccia, about 30 feet thick; some of the more sandy blocks are 3 feet across; they are angular and lie disposed in all directions in the less sandy Mallett dolomite.

The ridges of Mallett continue eastward for about one-third of a mile to lower, flatter farmed land, beneath which lies Parker slate, and the width of the farmed land is enough to give the regulation thickness of about 200 feet to this formation. Then, another small ridge exposes blue-gray dolomite with quartzites, from 20 to 25 feet thick and evidently the equivalent of the dolomite at the top of the Parker Cobble (Fig. 2, C, bed 5). What lies farther east is obscured by farmed land.

The dolomite at the top of the Parker formation is also exposed on the northern side of the road to St. Albans Bay, about 1.5 miles west of St. Albans. When the

[53] W. E. Logan: *Geology of Canada* (1863) p. 281.
[54] C. D. Walcott: *op. cit.*, p. 251.

road-cutting was new, in 1924, this dolomite was overlain by some black slate, which almost certainly represented the Middle Cambrian (St. Albans formation). Lower, or west, are the following strata of the Parker formation:

Bed		Thickness (Feet)
7.	Intraformational flat-pebble dolomite conglomerate	5
6.	White, massive sandstone, dipping 11° E.	5
5.	Gray, massive, evenly bedded, sandy, conglomeratic dolomite, with an abundance of Parker slate inclusions, which weather out, leaving holes in the rusty-colored strata	10
4.	Typical Parker slate exposed in road ditch	5
3.	Dolomite breccia of angular pieces of dolomite and Parker slate, as much as 4 feet across, in a paste of round-grained sandy dolomite	15
2.	Parker slate, seen to the north beneath the road and among the bushes, and containing *Nisusia festinata, Kutorgina cingulata,* and *Hyolithes americanus*	5
1.	Farther west, all is covered ground without rock exposures until the junction of the Mallett with the Winooski is reached. This interval may conceal 150 feet or more of Parker slate.	

South of St. Albans Bay village.—About half a mile to the south of the previous section (Fig. 2, C), another traverse was made from St. Albans Bay southeast to the Rugg Brook road (Fig. 1, traverse XXIII). The village of St. Albans Bay is on low ground to the west of the Highgate Springs outcrops of the Western Sequence, extensive exposures of which may be seen on the two lake terraces to the west of the Winooski escarpment. The series here consists of light-blue massive limestones and yellowish-weathering dolomites, dipping 20°E. The actual thrust contact between the Winooski and the Ordovician is not exposed.

To the east, the Central Sequence begins in a cliff, 30 to 50 feet high, with the well-bedded, massive Winooski dolomite dipping east about 9 degrees. Farther east is 25 feet of Mallett-like dolomite, then more Winooski dolomite. Next is a cliff of the transition Winooski-Mallett, and, still farther east, the typical Mallett. The thickness of the entire Winooski-Mallett series is approximately 375 feet.

Farther east, a long exposure of Parker slate, dipping about 10° E., extends nearly to the Rugg Brook road. A short distance to the west of the road vertical cliffs expose about 30 feet of the youngest part of the Parker formation; the upper 20 feet are massive sandy dolomites, and the lower 10 feet, coarse soft sandstones, which weather down into a loose ocherous sand. These beds dip about 15°E. Beneath the dolomite is exposed about 50 feet of decidedly micaceous Parker slate, which is more or less nodular and weathers out into ocherous spots. The basal 25 feet of the Parker is massive and coarse-grained, contains much fine sand, and weathers out into "iron-rust" holes, 2 inches or more across, which give it the appearance of a conglomerate. In places in the lower portion there are also quartzite lentils, 5 to 8 feet thick. The actual contact with the Mallett is rarely seen. In this area the thickness of the Parker formation is approximately 300 feet.

Conner farm, southwest of St. Albans.—On the Rugg Brook road, less than a mile south of the St. Albans road, fossiliferous Middle Cambrian, with its basal Rugg Brook dolomite breccia, is exposed. Beneath the latter is a long exposure of the upper dolomite beds of the Parker formation, extending west across the road and up to the top of the hill, where the dolomite lies on Parker slate. On this hillside

the dolomites occur in a monocline, with the dip steepening from almost horizontal to about 10° E.; the thickness does not appear to exceed 30 to 40 feet.

Near the mouth of Rugg Brook, east of St. Albans Bay.—A fine sequence of Cambrian begins about 3.5 miles southwest of St. Albans, near the mouth of Rugg Brook, and extends east to St. Albans Hill (Fig. 1, traverse XXV). In this section the Winooski formation has a thickness of 385 feet, and the Mallett above it is 190 feet thick. The Parker beds are not well exposed, but appear to be 115 feet thick.

Burlington.—In the Phelps stone quarry in South Burlington, the Monkton is represented by a fine-grained red quartzite with some interbedded white or pink, thin-bedded layers. Many of the shale partings show rippling, sun-cracking, and rain-pitting.

In the city of St. Albans, some of the sidewalks are paved with slabs of the Monkton, probably derived from the Burlington quarries. These show rippling, mainly of the translation type, although some examples of the oscillation type are also present. More interesting are the pieces that show beach or wash marks made by the run-up of the waves, indicating a flat beach. Half a dozen slabs were seen that were completely pitted by rain imprints.

Jones Hill.—The upper Monkton is well exposed along the roadside at Jones Hill, less than 12 miles south of Burlington, where the well-bedded brick-red quartzites are interbedded with thin shale layers. Some of these layers show sun-cracking, rain-pitting, and shallow oscillation ripples, also fucoidal markings and faint trails of some crawling animal. At least one massive pink dolomite, 5 to 6 feet thick, is found here.

Taconic sequence, West Castleton.—West Castleton is in the area of the Taconic Mountains, and this is why the few fossils recovered thereabouts are like those of the eastern side of the Hudson Valley. The significance of the dissimilar Cambrian faunas in northwestern Vermont and in the Taconic Mountains has already been discussed. According to Keith:

"The sequence consists almost wholly of slates, but it has also one thin formation of limestone and one of quartzite. By means of these two formations in the Lower Cambrian and one of red slate in the Ordovician, the order and structure of these formations can be disentangled. The limestone (Beebe) contains a good Lower Cambrian fauna, and one of the slates has Middle Ordovician fossils."[85]

The writer devoted a few days to the Castleton area with Keith in August 1921, looking for fossils in the purple and green, decidedly cleaved, roofing-slate belt along the western side of Bomoseen Lake. This quarry belt is no thicker than 50 feet, although in places as much as 150 feet of slate has been taken out. A short distance below the slate belt is a zone, 5 to 20 feet thick, of interbedded slate and thin gray limestone (1 to 2 inches), replete with calcite-filled tension cracks. Keith[86] has named this thin limestone belt Beebe, with the type area at Beebe Pond in Hubbardton, Vermont. It has fossils, and near West Castleton, in 1922, Dunbar and the writer collected from it many specimens of *Eodiscus speciosus, Hyolithellus micans,* and *Obolella crassa,* indicating a Lower Cambrian age equivalent to that represented in the vicinity of Troy, New York.

[85] Arthur Keith: *Stratigraphy and structure of northwestern Vermont,* Washington Acad. Sci., Jour., vol. 22 (1932) p. 399.

[86] *Op. cit.,* p. 402.

Beneath the quarry zone are isolated beds of dolomite, one of which is 2 feet thick and of wide distribution. The limestone and dolomite have here, as in northern Vermont, a certain amount of well-rounded sand grains. The Lower Cambrian of the Castleton area may be 3000 feet thick, and the bulk of it is composed of greenish slates.

At Cedar Point on Bomoseen Lake, southeast of West Castleton, another slate quarry has been cut in a closed and overturned syncline, and, here, the same thin-bedded Beebe limestone yielded a single specimen of *Eodiscus speciosus*.

A short distance to the south of Glen Lake, about a mile southwest of West Castleton, a small vertical cliff (jointed face) shows strata in a synclinal fold. Not far to the west of this cliff, beside the road, black slates interbedded with dark limestones are exposed, in which cleavage and bedding are alike. Faint impressions of trilobites can be seen in these beds, and, after considerable hunting, enough were found to indicate the head of an *Olenellus* and a poor specimen of *Eodiscus*.

The same folded Lower Cambrian slate belt is found to the east of Bomoseen Lake, and, near Twin Lake, thin-bedded limestones contain fragmentary trilobites. Species of *Protospongia* and pieces of *Olenellus* were seen at this locality.

MIDDLE CAMBRIAN

GENERAL STATEMENT

Paleontologists had long been looking for the Middle Cambrian in northwestern Vermont, but had failed to find evidence of such faunas anywhere throughout the St. Lawrence geosyncline. The surprise was great, therefore, when a local collector, G. E. Edson, of St. Albans, found such forms in the outskirts of that city, and still greater when Walcott determined the fossils to be of the Atlantic realm and specifically of the Acadian geosyncline. How this Atlantic sea got into northwestern Vermont around the New Brunswick geanticline remains an unsolved puzzle.

A detailed history of the finding of this Middle Cambrian, and, in fact, of the entire American Cambrian, has been given by Howell.[57]

The Middle Cambrian of Vermont consists of the St. Albans slate, with the Rugg Brook conglomerate at its base. The thickness of the slate is between 60 and 270 feet.

FORMATIONS AND FAUNAS

Rugg Brook Conglomerate.—The St. Albans slate has the Rugg Brook conglomerate at its base in three places. These localities are: Rugg Brook, the Conner farm (which shows the best exposure), and west of Rockledge. This conglomerate weathers a salmon color, and is up to 20 feet thick; some of the dolomite blocks in it are 4.5 feet across; they appear to be derived from the Parker formation. The paste is the usual round-grained quartz sand cemented by dolomite. The upper limit of the St. Albans is easily

[57] B. F. Howell: *The Cambrian Paradoxides beds of northwestern Vermont*, Vt. State Geol., 16th Bienn. Rept., 1927-1928 (1929) p. 249-273.

ascertained locally where the Mill River limestone conglomerate is present, for the latter is the basal member of the Upper Cambrian (Hungerford slate). Where this conglomerate is absent, it is difficult to determine the upper limit of the St. Albans, but generally the Hungerford is decidedly banded and it commonly has an obscure "horn-like" guide fossil (*Bovicornellum;* Fig. 10)[58] in the lower 100 feet, by which it can be differentiated from the other slate series. No fossils are known from the Rugg Brook conglomerate. For further discussion of this conglomerate, the reader is referred to a paper by Howell.[59]

FIGURE 10.—*Bovicornellum vermontense Howell*

Sketch of the horn-like fossil characteristic of the middle Hungerford slate. x6. After Howell (1934).

St. Albans Slate.—The type locality for the St. Albans formation is Adams pasture, on the western outskirts of the city of St. Albans (Fig. 1, traverses II and III). It was for the strata of this place and the wider area around St. Albans that Marcou[60] founded his St. Albans group, a term to which Walcott[61] called attention in 1891 and which he used in 1910. As is now plain, the term embraced not only Middle Cambrian but Upper Cambrian (Hungerford) as well. Howell[62] in 1929 restricted the name to the Middle Cambrian portion of the original definition, and gave a sketch map of the known distribution.

Previous to 1908, Edson had found Middle Cambrian fossils in Adams pasture, both in the Upper Cambrian limestone conglomerate (Mill River) and in loose blocks of the St. Albans slate on either side of the conglomerate. State Geologist George H. Perkins, learning of these specimens, borrowed them to show to Walcott, who pronounced them to be of the *Paradoxides* fauna. This discovery was announced by Perkins[63] in his biennial report for 1907-1908. Keith, in his preliminary mapping of the area in 1923, did not know what to do with this find, not being aware that, in the summer of 1922, Howell had re-located Edson's locality and found more than a hundred specimens of

[58] B. F. Howell: *Bovicornellum vermontense, a peculiar new Cambrian fossil from Vermont,* Wagner Free Inst. Sci., Bull., vol. 9 (1934) p. 112-113.

[59] B. F. Howell: *Cambrian Rugg Brook formation of northwestern Vermont,* Geol. Soc. Am., Pr. 1933 (1934) p. 334.

[60] Jules Marcou: *The Taconic and Lower Silurian rocks of Vermont and Canada,* Boston Soc. Nat. Hist., Pr. 1861 (1862) p. 241, 246.

[61] C. D. Walcott: *Correlation papers; Cambrian,* U. S. Geol. Surv., Bull. 81 (1891) p. 281.

[62] B. F. Howell: *The Cambrian Paradoxides beds of northwestern Vermont,* Vt. State Geol., 16th Bienn. Rept., 1927-1928 (1929) p. 265.

[63] G. H. Perkins: *Preliminary report on the geology of Franklin County,* Vt. State Geol., 6th Rept., 1907-1908 (1908) p. 209.

this Middle Cambrian fauna, most of them *in situ*. In September 1924, Keith, Prindle, and the writer found two other localities where Middle Cambrian fossils are in place, the first of these about 1.5 miles southwest of St. Albans, on Martin Conner's land (Fig. 1, loc. XXIV), and the other, 4 miles northeast of the city, near the Rockledge estate (Fig. 1, traverse XXII). This, then, fixed the presence of Middle Cambrian beneath the Upper Cambrian slate (Hungerford) in northwestern Vermont.

The St. Albans formation consists almost wholly of blue-black, sandy, micaceous slate; rarely, a bed is very limy, making impure limestone, and there are also thin layers and lenses of sandstone and a few dolomite beds as thick as 6 feet. In places, the St. Albans slates are banded, but less so than the younger Hungerford and Highgate ones. If unfossiliferous, the St. Albans is hard to distinguish from the Parker slate. The greatest thickness, 270 feet, is in Adams pasture (Fig. 1, loc. III); 3.5 miles to the south, all the Middle Cambrian is absent. To the northwest of Rockledge the thickness is about 60 feet, and in the next half mile to the north the formation vanishes. These variable thicknesses appear to be due to the erosion interval following the deposition of the St. Albans slate and before the overlap of the Hungerford slate. The accompanying map (Fig. 1) shows the distribution of the formation.

The St. Albans fauna has more than 50 new species and is characterized by the trilobites *Elyx* and *Centropleura vermontensis*, the latter of which is related to *Paradoxides* and is found elsewhere only in southeastern Newfoundland and in Sweden. Both are described by Howell.[64] The common fossils are Agnostids and *Lingulella*. The fauna as a whole marks the time as latest Middle Cambrian; it is described by Howell in a paper to appear in the August issue of the BULLETIN.

From the facts stated, it is apparent that a long time break between the Lower and the Middle Cambrian resulted from lack of deposition, and that this interval represents a loss of record, the length of which is equivalent to the early and the middle portions of the Middle Cambrian. The break between the Middle and the Upper Cambrian is much shorter.

DETAILED DESCRIPTION OF LOCALITIES

Adams pasture, west of St. Albans.—This is the type area for the St. Albans slate (Fig. 1, traverse III). The chief horizon for fossils is near the top of the formation, beneath the Mill River conglomerate at the base of the Upper Cambrian; another fossiliferous zone, far less rich than the other, is exposed about 70 feet lower, stratigraphically. The formation here has a thickness of about 270 feet between the Mill River conglomerate and the older dolomite horizon at the top of the Lower Cambrian (the distance on the road is 1585 feet, and the average dip is 10 degrees); exposures of the slate are, however, to be seen only near the top of the St. Albans formation.

[64] B. F. Howell: *Two new Cambrian trilobites from Vermont*, Wagner Free Inst. Sci., Tr., vol. 7 (1932) p. 4-8.

Two and a half miles north of St. Albans.—On the east side of the Swanton road less than 2.5 miles north of St. Albans (Fig. 1, loc. XXVI), a small outcrop, not more than 10 feet thick, of the Mill River limestone conglomerate is exposed at the base of the Hungerford slate. Other similar conglomerates crop out in the field farther to the east. The Mill River conglomerate is underlain by shale, which seemingly should be the continuation of the St. Albans slate of Adams pasture and that to the north. Along the Swanton road, the outcrops of this slate belt are limited.

Rockledge.—Rockledge is an estate, 4.5 miles north of St. Albans, on the road to Highgate Falls, and, here, the largest and most conspicuous marble lentil (bioherm) of the Hungerford formation is exposed (Fig. 1, loc. XIII) (page 1064). To the south of the estate, an east-west road leads to the Fonda quarry on the Swanton road (Fig. 1, traverse XXII). The highway to the west begins on Upper Cambrian (Georgia slate), and the basal Rockledge limestone conglomerate of this formation is to be seen shortly to the west and south of the road and barns. Beneath the Rockledge and for about 1250 feet west, the highly banded dark slate of the Hungerford formation dips east about 10 degrees, and in it, Howell has collected the horn-like guide fossil characteristic of that formation. The base of the Hungerford is indicated by small exposures of the Mill River limestone conglomerate. The thickness of the Hungerford slate is, therefore, about 215 feet, an unusually small figure, seemingly due to subsequent erosion at the top of the formation.

The Mill River limestone conglomerate at the base of the Hungerford slate shows in two small outcrops north of the road to the west. The conglomerate ranges in thickness up to 8 feet, and, as elsewhere, the limestone pieces of this member are mostly fragments of the Lower Cambrian lentils, and a few pieces of the Parker dolomite.

Beneath the Mill River conglomerate and the erosional unconformity lies the St. Albans slate, which extends west for 360 feet; below that is a fine exposure of massive dolomite and dolomite conglomerate, not unlike the Rugg Brook dolomite, from 10 to 25 feet thick and dipping 10° E., with thin-bedded strata at the base. This gives a thickness between 60 and 70 feet for the Middle Cambrian at this place; half a mile to the north, it is absent, and it is not seen elsewhere in that direction. The material of the Rugg Brook conglomerate looks like that of any of the older dolomites of the Parker slate, and can be distinguished only by its blocky and fragmentary nature; the conglomerate here is interpreted as Rugg Brook, lying upon a Lower Cambrian dolomite of the Parker formation, from which it has been derived, and over which it has been deposited. Farther west, the Parker slate includes two other local lenses of dolomite conglomerate.

Numerous fossils have been collected in the St. Albans slate west of the Rockledge estate, immediately beneath the Mill River conglomerate, the dark micaceous slate here yielding *Lingulella vermontana, Spinagnostus franklinensis, Champlainia rectimargo, Micragraulos decorus, Solenopleura franklinensis, Elrathia longa, Centropleura vermontensis,* and *Ptychoparia* (?) *convexa.*

South of the east-west road to Fonda quarry already mentioned, the whole of the St. Albans formation is well exposed, and Howell has collected many trilobites from it. The basal Rugg Brook dolomite member at this locality is, likewise, either a massive sandy dolomite or a conglomerate. It occurs in discontinuous outcrops, with a thickness varying between 10 and 20 feet; in one place, it has a lens of white quartzite, 30 inches thick, composed of rounded sand grains. Ten

feet above the Rugg Brook dolomite is exposed another sandy dolomite conglomerate, 10 feet thick, with blocks as much as 14 inches across. Beneath the Rugg Brook conglomerate lies the Parker slate.

Less than a mile south of the Rockledge outcrop of the St. Albans slate, the writer collected *Spinagnostus franklinensis*, which shows the presence of the Middle Cambrian.

Conner farm, southwest of St. Albans.—On the road to Rugg Brook, about a mile south of the St. Albans Bay road, and to the west of Martin Conner's home (Fig. 1, loc. XXIV), the Parker formation is exposed (p. 1037); it extends east of the road for about 150 yards. The base of the Middle Cambrian begins with the best exposure of the Rugg Brook conglomerate so far seen, composed of angular pieces of the Parker dolomite, in blocks as much as 4.5 feet across, embedded in a matrix of sandy dolomite. The conglomerate dips 15° E. and is about 15 feet thick. For 150 yards farther east, all the rocks are St. Albans slate (= 115 feet thick); beyond, they are covered by Champlain clays. The St. Albans here is a dark-blue slate with impure limestone bands. From it the writer collected *Lingulella vermontana, Spinagnostus franklinensis, Grandagnostus vermontensis, Conoides edsoni, Micragraulos decorus, Solenopleura franklinensis, Elrathia longa, Centropleura vermontensis,* and *Ptychoparia* (?) *convexa.*

Near the mouth of Rugg Brook.—Less than two miles south of Martin Conner's farm, another cross-road runs east to St. Albans Hill and west to the shore of St. Albans Bay, and along this road the total thickness of the Middle Cambrian can be measured (Fig. 1, traverse XXV). The formations are as follows:

Beginning at the west, the Winooski is overthrust upon the Highgate Springs marble of the Western Sequence, and extends east for half a mile, with a dip of about 10° E. The thickness is 385 feet.

Above the Winooski, in conformable relation, lies the Mallett, which continues east across the north-south road and Rugg Brook. On the east bank of the stream, a little cliff exposes a coarse intraformational dolomite conglomerate about 20 feet thick, with rare pieces of the Winooski formation, the whole bound together by a dolomite paste full of round-grain sand. Schuchert[65] originally mistook the position of this dolomite conglomerate and gave it the name of Rugg Brook, thinking that it lay at the base of the Middle Cambrian, but, during a visit to the locality in 1933, Howell proved to him that it is an intraformational conglomerate within the Mallett formation. Hence, the Mallett is now believed to continue from the brook east for about 350 feet, making the thickness of the formation here only 190 feet. The reason for this unusually low figure is unknown.

The Parker formation is believed to begin east of a 5-foot outcrop of dolomite along the roadside, above which slate continues east for 225 yards, indicating a thickness of about 115 feet. No fossils have been found.

In probable unconformity upon the Parker formation lies the St. Albans, which begins with a gray sandy dolomite, 10 feet thick, weathering a rusty red; to this basal member is now transferred the term Rugg Brook dolomite, for it, too, lies in the small valley of Rugg Brook. It is overlain by slates, which continue eastward for 400 feet, with a dip of 10° E., indicating a thickness of about 70 feet, instead of 250 feet as stated by Schuchert in 1933. Accordingly, the St. Albans and the Rugg Brook have a united thickness of 80 feet. The top of the

[65] Charles Schuchert: *Cambrian and Ordovician stratigraphy of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 25 (1933) p. 366.

St. Albans is determined by the presence of the Mill River conglomerate at the base of the Hungerford formation, which is well exposed in a west-facing cliff a few hundred feet south of the east-west road. No Middle Cambrian fossils have been collected here, but many were found by Howell a short distance to the south, in the Mill River bottom.

Fifty feet above the Mill River conglomerate, Howell found specimens of the guide fossil *Bovicornellum*, so characteristic of the lower Hungerford slate. These slates continue east with a low dip for nearly three-fourths of a mile, and are then overlain by the Rockledge limestone conglomerate.

Mill River.—The bottom land of Mill River, lying about a third of a mile to the south of the Rugg Brook locality just described, is the type locality for the Mill River limestone conglomerate of Howell,[66] which forms the base of the Hungerford slate. It is, in general, less than 8 feet thick, but in one place it swells to 15 feet, and all the limestone pieces and blocks are from the lentils derived out of the Parker slate. None longer than 5 feet was seen, and, only rarely, large pieces of Mallett dolomite (3 feet across) are exposed.

Beneath the Mill River conglomerate lie good outcrops of the St. Albans slate with fossils, but a short distance to the west these are covered by river wash. Howell collected many Middle Cambrian species at this locality.

UPPER CAMBRIAN

GENERAL STATEMENT

The presence of Upper Cambrian in northwestern Vermont was suspected as early as 1861. It was not until 1886, however, that Walcott found a few fossils, thought to be of this age, near Highgate Center, but even then the evidence was not fully convincing. In 1921, Keith found more fossils at this same place and in the same formation—the lower Highgate limestone in the village of Highgate Center (Fig. 1, loc. XXVII)—and Walcott again pronounced them to be of Upper Cambrian time. In consequence of Raymond's detailed studies of the trilobites (to be published in the August issue of the BULLETIN), however, it is now established that the Highgate formation is of the lowest Ordovician, and the equivalent of the Ceratopyge limestone of northwestern Europe. This conclusion is discussed on later pages.

On July 5, 1922, Keith, Sayles, and the writer found in the Highgate gorge (Fig. 1, loc. XI) a series of thin-bedded dolomites, some of which are limestones with an abundance of fragmented dwarf trilobites, and these fossils, when identified, demonstrated for the first time the actual presence of a varied late Cambrian fauna. The strata containing these trilobites are now called the Gorge formation. In subsequent years, Upper Cambrian fossils were found by Ulrich in the Milton dolomite, a formation that had been thought to be of Lower Cambrian age. From time to time, a rare trilobite was found in slates in the area south of High-

[66] B. F. Howell: *Base of the Upper Cambrian in northwestern Vermont*, Geol. Soc. Am., Pr. 1933 (1934) p. 336-337.

gate Falls, and these were thought to be correlates of the Highgate formation in the Highgate gorge and to the north. However, when Howell found numerous trilobites in the exposures of these slates on Hungerford Brook, south of Highgate (Fig. 1, loc. XXVIII), it became clear to him that the slates were older than the early Ordovician Highgate formation and, therefore, must be Upper Cambrian. Accordingly, in this paper, these strata are christened the Hungerford formation. Overlying them is a limestone breccia, first called Swanton by Keith and later re-named Corliss, which here and there has yielded a few fossiliferous pebbles containing trilobites, thought at that time to be of Lower Cambrian kinds. Because of its position, this so-called "Corliss" breccia was believed to be at the base of the Lower Ordovician. After long search, the writer secured many fragmented trilobites, and a number of these also were found in the paste of the formation, showing that they all belong to one fauna and that the "conglomerate" was not composed of foreign rocks but was a broken-up mass of local reefs of thin-bedded limestones and algoid bioherms, cemented together by calcium carbonate and abundant round-grain quartz sand. Raymond devoted a year of study to these trilobites, with the result that they are seen to be of early Upper Cambrian age. With this information in hand, it was apparent that the name Corliss could not be applied to all these widespread breccias, as the true Corliss has unmistakable Lower Ordovician fossils. These "Corliss" breccias are, therefore, here re-named Rockledge (Fig. 1, loc. XIII), and, inasmuch as they continue conformably and unbroken into the "Georgia formation" of Keith, at least the lower part of this thick series must also be of Upper Cambrian age.

It is now evident that there is much Upper Cambrian in northwestern Vermont, more than 700 feet in the Highgate Falls area and more than 1300 feet to the south in Georgia and Milton townships. The Upper Cambrian is, therefore, more varied faunally and lithologically than either the Middle Cambrian or the Lower Ordovician. Vermont has, indeed, a fine development of the Upper Cambrian, which will make it the standard area in the correlation of these formations throughout the St. Lawrence geosyncline.

The Upper Cambrian of northwestern Vermont occurs in eight formations, which appear to represent five distinct time intervals. (Table 1.) The oldest of these is (1) the Milton dolomite. It is overlain, apparently without break, by (2) the Hungerford slate, the formation of widest distribution, which has at its base locally (3) the Mill River limestone breccia. The Hungerford is overlain disconformably by (4) the widely distributed Rockledge limestone breccia, which goes unbroken into (5) the Georgia formation as restricted by Keith.

To the north, in the vicinity of Highgate, the Upper Cambrian sequence begins with (6) "Milton" dolomite, which is overlain by (7) "Hungerford" slate; these two formations appear to correlate with the true Milton dolomite and the Hungerford slate. Then there is a short break or thrust plane in the sequence, separating the dolomite and slate from the overlying (8) Gorge formation, which may correlate with the Rockledge and the Georgia formations.

These eight divisions will be described in detail.

FORMATIONS AND FAUNAS

Milton Dolomite.—The typical area for the Milton dolomite[67] is in Milton township, about 3.5 miles southwest of Milton village, or 2 miles northeast of Malletts Bay. According to Keith, the formation consists almost entirely of massive dolomite. It lies disconformably upon the Lower Cambrian Parker slate, and the two formations are separated by an erosion interval of long duration. Keith states that the Milton varies from dark bluish-gray or steel-gray through light gray to buff, and that beds of sandy dolomite are fairly common. It has much scattered chert, which stands out on weathered surfaces, together with "white quartz knots, or geodes with quartz crystals," and it includes also thin local wavy layers of *Cryptozoon.* Another peculiarity is its large content of scattered angular dolomite and dolomitic sandstone pebbles. The greatest thickness of the formation, according to Keith, is 700 feet, but the writer thinks the amount may be much less.

In the original account of the Milton, Keith regarded it as of Upper Cambrian age, although he found no fossils in the formation; later, however, he referred it to the Lower Cambrian. Ulrich, during the summers of 1930 and 1931, was the first to find fossils in the typical area, and they confirmed Keith's original age reference.[68] In his later field work, Keith continued his mapping north to the Canadian border, and, finding so many similar dolomites, confused some of them, and during his field season of 1932 also applied the term "Milton dolomite" to a much thinner one (about 30 feet thick) at the top of the Lower Cambrian (Parker). All these more northerly Lower Cambrian "Miltons" must now be excluded, and the term be restricted to the original meaning—namely, to the Upper Cambrian dolomite southwest of Milton, and to its equivalents.

The Milton fauna, as collected by the writer, consists mainly of brachiopods, among them *Lingulella acuminata, Syntrophia ? calcifera,* and *Huenella* cf. *H. battis.* It contains the gastropod, *Sinuopea,* and trilobites are represented by at least *Richardsonella.* As the Milton dolomite

[67] Arthur Keith: *Cambrian succession of northwestern Vermont,* Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 112-114.

[68] E. O. Ulrich: *Personal communication.*

is overlain by the Hungerford slate, a formation yielding early Upper Cambrian trilobites, the Milton should be of the older Upper Cambrian.

Mill River Limestone Breccia.—At only three places have local breccias been found at the base of the Hungerford shale, resting on the Middle Cambrian (St. Albans slate). At the type locality, 3.5 miles southwest of St. Albans (Fig. 1, traverse XXV), the Mill River breccia makes a low bluff to the east of Mill River, and attains a thickness of 15 feet, but, in places, it is less than 5 feet thick. Another easily accessible outcrop, also resting on the Middle Cambrian, is in Adams pasture on the outskirts of St. Albans; a third occurrence, with the same sequence, is a short distance to the west of the Rockledge estate, 4.5 miles north of the city (p. 1064).

In general, the Mill River breccia resembles the younger Rockledge breccia, but a close examination shows that the former has little of the slabby, dark-blue, fossiliferous limestone like that of the Gorge formation, pieces of which are so characteristic a feature of the Rockledge. It is, on the contrary, composed almost wholly of angular fragments of the bluish white limestone lentils that formed in the Lower Cambrian (Parker) slate. Some of these derived blocks are 20 feet across. Evidently, most of the Mill River breccia was formed where the invading sea encountered Lower Cambrian lentils. It also includes, rarely, small angular pieces, some of which are 8 inches long, of dolomite derived from the Parker or the Mallett formations, and, in Adams pasture, some of the limestone pieces are from the St. Albans slate. Locally, the Mill River has a basal white sandstone made up of re-worked, wind-blown sand.[69]

Edson collected the following fossils from the derived limestone pieces in the Mill River breccia, chiefly in Adams pasture: of Lower Cambrian age, *Lingulella franklinensis, Nisusia festinata,* and *Salterella pulchella;* of Middle Cambrian age, *Centropleura, Elyx,* and *Agnostus.* Other fossils, according to Howell, are: *Obolus matinalis, Huenella vermontana,* and *H. billingsi.*

Hungerford Slate.—The thick Hungerford slate series has until now been held to be the same as the Highgate formation, but, inasmuch as the fossils of the latter deposits are of early Ordovician times, whereas those of the Hungerford are of the older Upper Cambrian, the two are here distinguished for the first time as independent formations. The Highgate has thin-bedded impure limestones below, overlain by slates much like those of the Hungerford; the latter, however, is almost wholly of slate, with many white limestone bioherms and a few lentils of dolo-

[69] B. F. Howell: *Base of the Upper Cambrian in northwestern Vermont,* Geol. Soc. Am., Pr. 1933 (1934) p. 336-337.

mite. The Highgate is known only from Highgate Falls northward into Canada. The "Hungerford" crops out northwest of Highgate Center and, more typically, from Hungerford Brook south "to the latitude of Burlington, where it is cut off by the Hinesburg overthrust." [70] The name is taken from Hungerford Brook,[71] 1.25 miles south of Highgate (Fig. 1, loc. XXVIII), the type area being where the brook crosses the Highgate road, and where most of the known (but as yet undescribed) fossils have been collected by Howell.

The Hungerford formation is composed almost wholly of blue-black to gray, well-banded slates, with here and there local lentils of yellowish- or reddish-weathering, sandy dolomites, varying in thickness from 1 to 10 feet. The complete development is seen west of Georgia Center, where the thickness is about 700 feet, and the same thickness appears to be present in Milton township. The Hungerford slate, as a rule, overlies the Mill River breccia, and where the latter is absent, the slate rests either on the Middle Cambrian (St. Albans) or on the Lower Cambrian (Parker). It is overlain disconformably by the Rockledge breccia or by the Georgia formation.

The guide fossil to the lower 100 feet of the Hungerford formation is *Bovicornellum vermontense* Howell (Fig. 10), a form whose biogenetic relations are unknown, although it is common. Nothing of the original shell of these tiny fossils is present; only casts of the outer form are found. These are slightly twisted, horn-like, smooth holes in the slate. Undescribed brachiopods and rare trilobites are found with *Bovicornellum*. A simple type of *Agnostus*, still undescribed, has been collected in higher strata at several localities, and a species of *Lingulella* is more often seen.

In a lentil of limestone in the middle part of the Hungerford slate, 4 miles southwest of St. Albans, the writer found two trilobites, which have been described by Raymond as *Hemirhodon schucherti* and *H. viator*.

As Howell has long been collecting the fossils of the Hungerford slate, the writer asked him what age he assigned to these strata. His reply was:

> "The trilobites of the *Bovicornellum* beds and the beds near the bridge over Hungerford Brook are somewhat similar, but do not appear to be identical. Both of these faunas are, I think, of Upper Cambrian time (and probably not greatly different in age), and both are probably older than the Gorge faunas. . . . One of the Agnostians in the Hungerford Brook faunas is *Pseudagnostus*, an Upper Cambrian genus."[72]

The Hungerford formation is famous for its many limestone bioherms, in which lentils grew in an environment of mud during the time when

[70] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 377.

[71] Edward Hungerford was one of James Hall's assistants, and a lifelong friend; he later became professor of geology at the University of Vermont.

[72] *Personal communication.*

FIGURE 1. LIMESTONE REEF (BIOHERM)
Near view (Fig. 1, loc. XII). This reef was made by algae growing in vertical prisms. Photograph by Arthur Keith.

FIGURE 2. LARGE BIOHERM
One of the largest bioherms in Vermont; it is at the top of the Hungerford slate and was once buried in the Rockledge breccia. Rockledge estate, 4 miles north of St. Albans (Fig. 1, loc. XIII). Photograph by R. W. Sayles.

VERMONT BIOHERMS

FIGURE 1. ROCKLEDGE BRECCIA
Vertical cliff, 13 feet high, made up of basal part of Rockledge breccia, 3 miles north of St. Albans (Fig. 1, loc. XIII). Basal bed (3 feet thick) shows foresetting, and is followed upward by a 4-foot zone, mostly of sandstone, with fragments of trilobites and a few pieces of limestone; higher, the breccia becomes more and more marked. Photograph by A. C. Swinnerton.

FIGURE 2. BASAL HIGHGATE THRUST BRECCIA
Showing how this breccia is thrust into underlying upper Gorge limestones. Highgate gorge (Fig. 1, loc. XI). Note brecciation of thin beds around limestone block on which hammer rests. Photograph by A. C. Swinnerton.

this formation was being deposited. They are found throughout the formation. These white to gray marble-like masses resist erosion more than does the slate, and, weathering out, form prominent points in the landscape (Pl. 5). Their maximum size is "about 200 feet long and 80 feet wide, and they project from 10 to 20 feet above the general level. . . . They occur in clusters as well as individually." [73]

These lentils are homogeneous masses of dense limestone, which show no bedding. They appear to be small reefs, largely or wholly of algal growth. Cumings has termed such growths bioherms.[74]

Blocks and small pieces of these lentils, or of similar ones derived from the Lower Cambrian (Parker) slate, are seen in all the Mill River, Rockledge, and Corliss breccias. These detached masses have apparently been moved short distances only, but none of the large lentils has been shoved about by the waves of the sea. They are where the sea found them, but before the marine invasion that buried them took place, they may, during land intervals, have slid down slopes, so that some of them now lie unconformably upon strata other than those in which they grew.

Rockledge Limestone Breccia—At, or near, the base of the Georgia formation are the local reefs of the Rockledge breccia (Pl. 1; Pl. 5, fig. 2), heretofore called Swanton and, later, Corliss conglomerate by Keith.[75] Since 1932, it has become known that the breccia at the type locality of Corliss ledge (Fig. 1, loc. XXXIV) has unmistakable Ordovician fossils, whereas the so-called "Corliss" elsewhere has early Upper Cambrian trilobites. For these older breccias, a new name is therefore needed, and Rockledge is selected, after an estate on the Highgate road, 4.5 miles north of St. Albans (Fig. 1, east end of traverse VIII).

The Rockledge formation extends with marked interruptions from a point a mile south of the Missisquoi River to a point 2 miles south of Georgia Plains, a distance of 13 miles. Where the breccia is absent, the Georgia slate rests disconformably on the Hungerford slate, the best of these contacts being 4 miles north of St. Albans, on the west side of the road back of the barns (Fig. 1, south of loc. XIII).

The Rockledge breccia is composed, in the main, of small pieces of blue slabby limestones similar to those of the upper Gorge beds. Pieces of sandy dolomite are also present, and these appear to have been derived from local dolomites in the higher Hungerford slate. Probably from one-eighth to one-fourth of the Rockledge consists of large angular pieces of Hungerford limestone lentils, up to 10 feet across, and, locally, the breccia

[73] Arthur Keith: *op. cit.*, p. 376.
[74] E. R. Cumings: *Reefs or bioherms*, Geol. Soc. Am., Bull., vol. 43 (1932) p. 337-338.
[75] Arthur Keith: *op. cit.*, p. 377-378.

has formed around an entire Hungerford lentil, as at Rockledge (Pl. 5, fig. 2), and again less than a mile south. The breccia is cemented by lime or dolomite paste, which is replete with round grains of etched sand. In places, so much sand is present as to make basal sandstones 10 feet thick, or conglomeratic sandy dolomites anywhere up to 10 feet thick. This sand appears to be of dune origin, and originally was washed up by the invading Lower Cambrian sea; later, it was weathered out of these formations and re-deposited in the Rockledge breccia and other formations. In the basal sandstones of the Rockledge have been seen bits of trilobites too small to name; in one of the interbedded sandstones of the breccias, however, Raymond collected two identifiable trilobites, and many species have been found in the sandy limestone pebbles or in strata *in situ* near the base of the formation. These facts also show that the Rockledge breccia is of marine origin.

The widest exposures of the Rockledge breccia are about half a mile west of Georgia Center (Fig. 1, loc. X), and it was this outcrop that Walcott [76] noted as his division 9 of the Georgia group.

Where the Rockledge breccia is present, there is no Gorge formation. The Rockledge lies disconformably on the Hungerford formation. This relationship holds true from the southern area of the Rockledge breccia north to a point near Highgate Falls. In other words, the Rockledge and the Georgia slate begin a new cycle of marine invasion.

In Table 2 are listed 29 species of the Rockledge fauna, of which 10 were found *in situ* and 23 in pebbles, 4 being found in both groups. Five of the species and seven of the genera are found also in the Gorge formation, and one species is known in the Hungerford slate.

Three and a half miles north of St. Albans, the matrix of the Rockledge breccia is sandy, in places made up largely of wind-blown quartz grains, the whole bound together by a calcareous cement. At, or near, the base are definite layers of the matrix, partly a calcareous sandstone and partly a sandy dolomite. These basal *in situ* layers have furnished ten species of trilobites, as shown in column 1 of the faunal table. Near the top of the breccia, in a layer of sandstone, also *in situ*, Raymond found *Coleopachys strix* and *Ucebia lata*. The fact that *Ucebia lata* occurs *in situ* and is common elsewhere in the pebbles, he says, "demonstrates that the basal layers are a part of the Rockledge itself. The presence of the same species and of *Hysteropleura macgerriglei* in the boulders in the breccia northwest of Georgia Center, and of *Coleopachys strix* in a pebble at Rockledge, indicates that some, at least, of the pebbles are penecontemporaneous with the matrix. This suggests that the conglomerate is of

[76] C. D. Walcott: *Second contribution to the studies on the Cambrian faunas of North America*, U. S. Geol. Surv., Bull. 30 (1886) p. 17.

reef origin, its materials subject to occasional emergences, partial consolidation, and destructive wave action.

"The presence of *Paracrepicephalus, Idahoia, Dunderbergia,* and *Illaenurus* in the matrix fixes the age of the Rockledge definitely as Upper

TABLE 2.—*Trilobites of the Rockledge Breccia*

Trilobite	In situ	Pebbles		Genera also in Upper Cambrian	In Hungerford formation	Genera also in Gorge formation	Species also in Gorge formation
	Loc. 1 3.5 miles north of St. Albans	Loc. 2 northwest of Georgia Center, near lentils	Loc. 3 Rockledge, 4.5 miles north of St. Albans				
Hysteropleura macgerriglei Raymond	x	x		x			
Paracrepicephalus cf. *P. thoosa* (Walcott)	x			x			
Idahoia minor Raymond	x			x			
Dunderbergia vermontana Raymond	x			x			
Ucebia simplex Raymond	x			x			
U. lata Raymond	x, x′	x	x	x			
Illaenurus breviceps Raymond	x	x	x			x	x
Illaenurus sp. indet.	x						
Coleopachys pisum Raymond	x						
C. striz Raymond	x′		x				
Cephalocoelia ovoides Raymond	x						
Onchonotus eminens Raymond		x				x	
Distaseris acuta Raymond		x					
Calvinella obsoleta		x		x		x	
Tsinania scrinium Raymond		x		x			
Steriocombus princeps Raymond		x					
Hemirhodon minor Raymond		x		x			
H. schucherti Raymond		x	x		x		
Lecanopleura interrupta Raymond		x				x	x
L. inflecta Raymond		x					
Agnostus innocens Clark			x			x	x
Protillaenus marginatus Raymond			x				
Stenopilus pronus Raymond			x			x	x
Kaninia ? *platys* Raymond			x				
Bynumia leptogaster Raymond			x	x			
Greylockia minuta Raymond			x				
Maryvillia triangularis Raymond			x	x		x	x
Homodictya imitatrix Raymond			x				
Stenelymus kobayashii Raymond			x	x			
Acrohybus argutus Raymond			x				

x—below.
x′—above.

Cambrian. If present ideas as to the vertical distribution of *Paracrepicephalus* are correct, it is a rather early Upper Cambrian fauna." [77]

Raymond thinks that the Rockledge conglomerate may be older than the upper part of the Gorge formation, and that it represents a shallow-

[77] P. E. Raymond: *Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont.* Geol. Soc. Am., Bull., vol. 48 (1937). In press.

water, off-shore reef deposit around bars and islands. His reason for this opinion is the general similarity of the fauna of the Rockledge and the Gorge formations; differences in the faunules he thinks are due to facies conditions—one set of organisms living in the agitated waters around upstanding reefs (Rockledge) and the other living in deeper quiet waters (Gorge). The round-grained, wind-borne sand in the Rockledge may be the residue of older Cambrian dolomites, nearly all of which have some similar sand, or it may have come from the Adirondack region, which supplied so much similar sand to the Potsdam formation. Accordingly, Raymond thinks it "very probable that the Rockledge is a series of such bioherms, all formed at one time. The action of the waves would tend to concentrate the rather coarse wind-blown sand in the shallows" around these reefs. He thinks, also, that later studies may show that the Rockledge fauna is of about the age of the Cedaria fauna in the upper part of the Eau Claire formation of Wisconsin.

Georgia Slate.—Keith [78] restricted the old broad term Georgia (1861) to this formation, citing as the typical area the eastern part of Georgia township, in the vicinity of Georgia Center. The formation here "forms a belt three miles wide, and its thickness is likely to be over rather than under 2000 feet." From here, it crops out nearly continuously to a point beyond the Canadian border. The Georgia slate is mostly a dark blue-gray, fine-grained, and more or less siliceous slate, which is strongly cleaved and folded. As a rule, it is not well banded, but otherwise, it resembles the Hungerford and Highgate slates, except that it is generally lighter in color and coarser in grain. It also has dolomitic sandy layers and thin bands of sandstone made up of rounded sand grains. No fossils have been found in the type area of the series.

As the Georgia slate is intimately tied stratigraphically with the Rockledge breccia, now known to be of early Upper Cambrian time, there can be no doubt that these slates are also of this epoch.

Gorge Formation.—This name, though inappropriate, seems to be the best that can be found to designate the strata beneath the Highgate formation in the Highgate gorge (Fig. 1, loc. XI), which have an estimated thickness of 220 feet. The detail of this formation is given under the heading *Highgate gorge,* at the end of the Upper Cambrian discussion. Beds of limestone at three different levels in the upper thin-bedded part of the Gorge formation abound in fragmented trilobites, indicating a late Upper Cambrian age. As these fossils are all different from those of the Milton dolomite, which appears to be a considerably older formation

[78] Arthur Keith: *Cambrian succession of northwestern Vermont,* Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 123.

farther south, the name Milton cannot be used for the strata in the gorge.

The Gorge formation is easily divided, on lithologic grounds, into two parts, as follows:

The upper Gorge (Pl. 2) consists of 145 feet of thin-bedded dolomite and limestone, black dolomitic slate, and at least eight intraformational flat-pebble breccia beds, made up of pieces of dolomite and limestone; the lower part has thick beds of sandstones. Keith [79] called this series of beds Missisquoi, but that term was later found to have been already applied by the Vermont Survey to a formation on the eastern side of the Green Mountains.

The upper Gorge is separated from the Highgate formation by a thrust plane, above which is a great breccia zone, the basal member of the Highgate (Pl. 6, Fig. 2). It will be described in connection with the Ordovician section at Highgate Falls.

The lower Gorge is made up of massive dolomite, more or less replete with thin-bedded black dolomite inclusions, making the whole series conglomeratic. No fossils have been seen. At the base is a massive bed consisting of water-worn dolomite pebbles with blue slate bands, 4 feet of which shows above river level. The entire division has a thickness of about 75 feet.

This same basal conglomerate is also exposed at a point a mile northwest (Fig. 1, east end of traverse V), in contact with a slate (= "Hungerford") carrying a graptolite (*Dictyonema*), which cannot be older than Upper Cambrian. The same contact may be inferred in the Highgate gorge.

Much trilobite material has been collected from the two closely associated, fossiliferous beds (zones 3 and 2) in the upper Gorge division and, according to Raymond, the revised list is as follows:

Agnostus innocens Clark
A. trisectus Salter
A. insuetus Raymond 1924
Pseudagnostus extumidus Raymond 1924
Peronopsis planulata Raymond 1924
Phalacroma parilis (Hall)
P. cyclostigma Raymond 1924
Stenopilus pronus Raymond 1924
S. brevis Raymond 1924
Hemirhodon juvenis (Raymond) =
Corynexochus juvenis Raymond 1924
Acheilus macrops Raymond 1924
Idiomesus tantillus Raymond 1924
Phoreotropis puteatus Raymond 1924
P. transversus Raymond 1924
Zacompsus clarki Raymond 1924
Onchonotus nasutus (Walcott)
Phylacterus saylesi Raymond 1924
P. fraternus Raymond 1924
Blountia imitator Raymond 1924
Maryvillia triangularis Raymond 1924
Lloydia seelyi (Walcott)
Plethometopus armatus (Billings)
P. latus (Raymond) 1924
P. laevis (Raymond) 1924
P. angustus (Raymond) 1924
P. convergens (Raymond) 1924
Apatokephaloides clivosus Raymond 1924

[79] Arthur Keith, *in* P. E. Raymond: *New Upper Cambrian and Lower Ordovician trilobites from Vermont*, Vt. State Geol., 14th Bien. Rept., 1923-1924 (1924) p. 137.

A. inflatus Raymond 1924
Protapatokephalus minutus (Raymond) = *Hungaia minuta* Raymond 1924
Saukia sp. ind. = *S. lodiensis* (Whitfield) Raymond 1924
Saukiella inops Raymond = *Saukia stosei* (Walcott) Raymond 1924
S. dunbari (Raymond) 1924 = *Saukia dunbari* Raymond 1924
Richardsonella germana Raymond 1924
Distazeris spicatus (Raymond) = *Acheilus spicatus* Raymond 1924
Illaenurus quadratus Hall
I. breviceps Raymond 1924
I. laevis Raymond 1924
Platycolpus dubius (Billings)
Ptychaspis affinis Raymond 1924
Keithia schucherti Raymond 1924
Ambolium lioderma Raymond 1924

Forty-one named trilobites in 27 genera are now known from the two higher fossiliferous zones. Of the species, 10 come from the lowest faunal assemblage (zone 1) of the Gorge formation, and 11 genera are common to the two upper divisions. Of the 27 genera, only 4 recur either in the Hungerford or in the Highgate formations—namely, *Agnostus, Hemirhodon, Maryvillia,* and *Lloydia*—and, of the species, only *Maryvillia triangularis* continues upward.

A number of other fossils have been collected from zone 2. These are *Lingulella* cf. *L. acuminata, Schizambon* cf. *S. manitouensis* Walcott, orthocerids, *Huenella* related to *H. simon,* and a gastropod.

About 80 feet lower in the upper Gorge, the following trilobites (zone 1), listed by Raymond, are found; those marked with an asterisk are also found in the two higher fossiliferous zones:

Pseudosalteria laevis Raymond
**Agnostus innocens* Clark
**Phalacroma parilis* (Hall)
**Phoreotropis puteata* Raymond
Tostonia duplicata Raymond
T. unisulcata Raymond
Liostracinoides vermontanus Raymond
Chariocephalus peloris Raymond
Diatemnus miculus Raymond
Aposolenopleura dunbari Raymond
Protapatokephalus arctostriatus Raymond
P. spiculatus Raymond
P. granulatus Raymond
**Apatokephaloides clivosus* Raymond
**A. inflatus* Raymond
**Acheilus macrops* Raymond
Dikelocephalus insolitus Raymond
Richardsonella iole (Walcott)
Highgatea laeviuscula Raymond
H. bisinuata Raymond
Calvinella sp. ind.
Stenosaukia irrasa Raymond
**Illaenurus breviceps* Raymond
I. punctatus Raymond
**Onchonotus nasutus* (Walcott)
Lecanopyge expansa Raymond
Heterocaryon platystigma Raymond
Bathyurellus ? antiquus Raymond
Plethopeltis incisa Raymond
**Plethometopus latus* Raymond
Strotocephala howelli Raymond
**Stenopilus pronus* Raymond
Platydiamesus inornatus (Raymond)
P. depressus Raymond
P. subtilis Raymond
Lecanopleura inflecta Raymond
Punctularia ornata Raymond

This lowest fauna of the upper Gorge (zone 1) has 37 forms in 29 genera, and, of the species, 10 are also found in the higher zones. Of the 29 genera, 13 are found also in the higher Gorge.

Raymond [80] says it is obvious that the trilobite faunules of the lowest zone (1) of the upper Gorge and the two higher zones are "rather closely related," and that the three zones "have enough species in common to indicate that no great amount of time had elapsed between the dates of the deposition of the younger and the older layers." *Tostonia, Liostracinoides, Chariocephalus, Saukia,* and *Illaenurus* are typical Upper Cambrian genera of the upper Mississippi Valley. The fauna of the Gorge formation thus indicates that these strata are of late Upper Cambrian time and are considerably younger than the Hungerford and the Rockledge formations. Accordingly, the Gorge formation may yet be found in the upper part of the Georgia formation.

"Hungerford" slate northwest of Highgate Center.—The sequence here is seen below the Gorge formation 1.25 miles northwest of Highgate Center (Fig. 1, east end of traverse V). The description proceeds from younger beds to older, because the youngest formation is the dated Gorge formation. The place is the small road-metal quarry (Fig. 11), approximately 150 feet along the face, on land belonging to Samuel C. Russell, a mile N 15° W of Highgate Center. For a long time, this formation was spoken of as the Russell slate, but, as this term is already in use, it is here abandoned in favor of "Hungerford" slate. The hill above the quarry shows massive, light-gray, conglomeratic dolomite, not more than 80 feet thick, down to the slate formation in the quarry. This overlying dolomite correlates with the nearby lower Gorge division, and agrees in lithology and stratigraphic position with the same formation in the Highgate gorge to the southeast.

When this quarry was first seen by the writer, many years ago, the opening was small, and minute brachiopods (*Acrothele* and *Acrotreta*), which appeared at the time to be of Lower Cambrian species, were collected in thin-bedded shaly dolomite; these dolomitic beds are now believed to be in the lowest part of the Gorge formation. Keith and the writer originally believed these underlying slates to be the Parker formation (Lower Cambrian) and so correlated them. Hence, the older dolomite to the west of the Russell quarry was considered to be Mallett, and a great break, representing all the Middle Cambrian, was placed above the slate and beneath the upper dolomite. Subsequently, the writer found a graptolite at the top of the "Hungerford" slate; this puzzled him until Howell suggested, on a visit to the quarry in 1933, that the graptolite-bearing bed might be Upper Cambrian, and, as the sequence was then

[80] P. E. Raymond: *Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont,* Geol. Soc. Am., Bull., vol. 48 (1937). In press.

better exposed through further quarrying, this suggestion was found to be correct.[81]

The slate formation which lies unconformably beneath the lower Gorge dolomite is here named the "Hungerford" slate, because it lies on a dolomite that appears to be the "Milton." The slate has a probable thickness

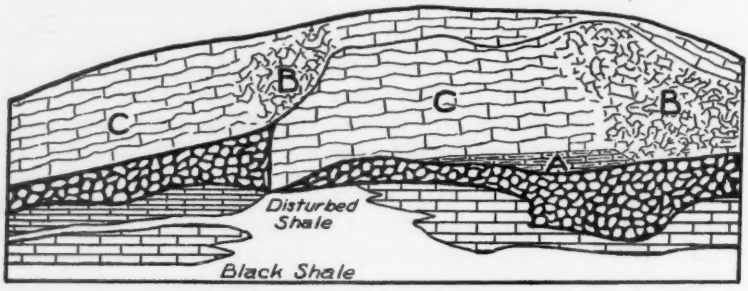


FIGURE 11.—*The Russell road-metal quarry*

Near Highgate Center (Fig. 1, east end of traverse V). At top, dolomite of lower division of the Gorge formation and its basal dolomite breccia thrust over, and lying disarranged upon, the "Hungerford" black slate of the Upper Cambrian. (B, B) Much broken, and (C, C) disturbed and fractured, Gorge dolomite. (A) Shaly dolomite with *Acrothele* and *Acrotreta*, formerly thought to indicate the Lower Cambrian. The black shale contains *Dictyonema schucherti* Ruedemann.

of less than 200 feet. Little of these soft beds is exposed anywhere in the vicinity, and the uppermost 6 feet of black slate in the quarry dips about 20 degrees to the east. These black slates have yielded rare specimens of *Dictyonema schucherti* Ruedemann.[82]

The Russell quarry seems to show a definite erosional unconformity between the lower Gorge dolomite and the "Hungerford" slate (Fig. 11). This is the view of all geologists who have seen the exposure, but how important this break is must be determined by the fossils found above and below it. As the "Hungerford" slate has only the single graptolite, and as these organisms are unknown in formations older than Upper Cambrian, the slate cannot be older than that epoch. That the younger Gorge dolomite is also of late Upper Cambrian age has been shown by the abundance of late Cambrian trilobites.

The writer has studied the Russell quarry many times during the past fifteen years, but not until 1932 was the opening large enough to show clearly the succession of strata and the nature of the contact between the Gorge dolomites and the older "Hungerford" slate.

At the base of the Gorge dolomites in the Russell quarry, there is a

[81] B. F. Howell: *Base of the Upper Cambrian in northwestern Vermont*, Geol. Soc. Am., Pr. 1933 (1934) p. 336-337.

[82] Rudolf Ruedemann: *The Cambrian of the upper Mississippi Valley, part 3, Graptolitoidea*, Pub. Mus. Milwaukee, Bull., vol. 12 (1933) p. 321.

conglomerate, irregular in thickness, of dark-blue pebbles of dolomite, which are rolled and rounded, in a sandy and muddy matrix, hardly any of the pieces being longer than 6 inches. Beneath this zone—the line of unconformity—and embedded in the older slate are disconnected slivers of massive dolomites several feet long, and, above the rolled pieces, all the dolomites are in a fractured and somewhat confused condition, so that the detail cannot be readily described or shown in photographs, but must be seen to be appreciated. Above the basal conglomerate, the dolomite higher up looks more like a disrupted mass of once regularly bedded, thin and massive layers. Furthermore, the conglomerate of water-worn dolomite pieces is exactly like the thicker, undisturbed bed at the base of the lower Gorge division at Highgate Falls, but there the contact with the older slate is hidden beneath the river.

"Milton" Dolomite.—The "Hungerford" slate makes a narrow valley through which runs the north-south road; to the east is a low ridge with the Gorge dolomite, and, to the west, a higher ridge made of light-gray sandy dolomite, dipping about 20 degrees to the east and having a probable thickness of 300 feet. Keith and others have long correlated this latter dolomite with the Mallett, but this correlation is questioned by Howell,[88] for the strata are underlain by fossiliferous Lower Cambrian (Parker) slate, a formation younger than the Mallett, already described.

The "Milton" sandy dolomite merges downward with gray, coarse-grained, thin-bedded sandstones, 10 to 15 feet thick, and dipping about 10°E. These thin-bedded strata make a narrow terrace in front of, and beneath, the westerly facing cliff of the light-gray massive dolomite already mentioned; beneath the terrace is the Parker formation (Fig. 1, western black square of traverse V). The contact between the thin and the massive dolomites is not exposed, but, as this dolomitic series is so very different in character from the underlying Parker, there can be no doubt about the dolomite and sandstone beds belonging together, and they are probably of Milton age.

At this point, attention should be directed to the fauna in the Rockledge breccia, already mentioned, which appears to indicate that an Upper Cambrian horizon, older than the Gorge formation, must be present in northwestern Vermont, and it seems probable that these fossiliferous pebbles in the Rockledge were derived from the basal "Milton" dolomitic sandstones.

DETAILED DESCRIPTION OF LOCALITIES

Cobble Hill.—This is the most southern of the Upper Cambrian localities (area south of map, Fig. 1), and the others occur progressively northward.

[88] B. F. Howell: *Base of the Upper Cambrian in northwestern Vermont*, Geol. Soc. Am., Pr. 1933 (1934) p. 336-337.

Ten miles north of Burlington, or 3.5 miles in a straight line southwest of Milton, and three-fourths of a mile southwest of the landmark, Cobble Hill (860 feet high), are large outcrops of the bluish gray Milton dolomite. It was here, just east of an abandoned road, that Ulrich, Bridge, and Ruedemann collected the first fossils known from the Milton formation; this is the type area for the formation.

The dolomite at this place is massive, with intraformational fragments of different kinds of dolomite, some of which are almost a sandstone composed of rounded sand grains. All weather light-gray and show an exceedingly jagged surface, which is studded with siliceous white veins, sandy lumps, and, rarely, with silicified fossils and wavy bands of *Cryptozoon*. The thickness seen exceeds 200 feet, but, as there is much more dolomite to the west of the abandoned road, it may well be that 700 feet of strata is present here, as stated by Keith.[84] The beds dip about 10° E., the dip increasing to 20 degrees farther east, until, finally, the dolomites are overlain by at least 500 feet of the banded Hungerford slate.

Siliceous fossils are best seen in beds that do not weather out much chert, and, from such beds, Dunbar and the writer collected *"Syntrophia" elax* Clark, *Huenella* cf. *H. battis* (rare), *Sinuopea*, and *Richardsonella* (identifications by G. A. Cooper). None of these species is present in the Gorge formation to the north.

To the east of the Milton dolomite, the Hungerford slate continues for three quarters of a mile, to Cobble Hill. This area was examined by Longwell, who reports that the slate is folded, though not closely folded. At the west, the bedding and the cleavage are the same—about 35° E.—but beneath Cobble Hill the cleavage rises to 70° E., and this difference in angle is thought to have been caused by the overthrusting of the dolomite. The cap rock of Cobble Hill is a blue-gray, yellow-weathering, massive Lower Cambrian dolomite, which Keith interprets as an outlier (*Klippe*), an overthrust part of the Eastern Sequence. The dolomite that makes Cobble Hill dips 40° E., and lies upon the Hungerford slate. Beneath Cobble Hill, on its eastern side, the Hungerford slate is also exposed just west of the Milton-Colchester public road, where it appears to be more than 500 feet thick.

Five miles directly north, at another landmark, known as Arrowhead Mountain, more Hungerford slate is exposed, and over it is thrust the Cambrian Brigham Hill graywacke of Jacobs;[85] the Brigham Hill is believed to continue north into Clark's Oak Hill slice,[86] which has Lower Cambrian fossils.

Allen Brook near Cobble Hill.—Ulrich discovered another place for Upper Cambrian fossils, 4.25 miles in a straight line southwest of Milton and a quarter of a mile east of Allen Brook. This locality was visited on July 17, 1932, by Longwell, Dunbar, and Schuchert, who saw here about 150 feet of the upper Milton, consisting of thick-bedded yellowish to gray dolomites, which are more or less sandy, with decidedly sandy beds and, in the upper part, sandstones up to 10 feet thick; there are also horizons of intraformational breccia having angular pieces of dolomite and sandstone as much as 18 inches across.

The structure is an arch, striking N 30°W, with the eastern, descending limb showing most of the thickness—at least 150 feet—and dipping 15° E. The western limb dips more steeply to the west (28 degrees). The bedding planes show much intraformational breccia; most of the pieces are small, but some are 2 feet across,

[84] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 372.

[85] E. C. Jacobs: *Green Mountains of northern Vermont*, Geol. Soc. Am., Pr. 1934 (1935) p. 85.

[86] T. H. Clark: *Structure and stratigraphy of southern Quebec*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 1-20; *Lower Cambrian series from southern Quebec*, Roy. Canad. Inst., Tr., vol. 21, pt. 1 (1936) p. 135-151.

indicating that waves often tore up the sea bottom. Near the top of the western limb, in a sandy dolomite, 6 inches thick, are abundant specimens of the long-ranging *Lingulella acuminata*. The sand grains are rounded and appear to be of wind-blown origin, but may have been derived, during erosion intervals, from older dolomites known to have much wind-blown sand. This fossil horizon appears to be about 100 feet beneath the top of the beds of the eastern limb.

Milton.—In 1921, Keith and the writer saw extensive outcrops of the Milton formation to the northwest of Milton, where it is a massive-bedded series of sandy dolomites with intraformational conglomerates. The fragments are variable in quantity and character, with most of the pieces angular and consisting of several kinds of sandy dolomites; there are many flat-pebble dolomite inclusions, with the pieces up to 6 inches long. From this, it is evident that thin beds of dolomite must have been present to have furnished these pebbles in the Milton dolomite.

Two miles south of Georgia Plains.—Wide exposures of the Hungerford slates are to be seen to the southeast of Georgia Plains and for 2 miles along the road south of that village. Two miles south of the village and a short distance northwest of a farmhouse, the road crosses a small exposure of the Rockledge breccia piled around two bioherms, one 15 feet across and the other larger, although neither appears to be more than 4 feet thick. The breccia lies disconformably upon the Hungerford slate. This locality is interesting mainly because the bedding and the cleavage in the Hungerford coincide, breaking out large sheets, which dip gently to the east. On a short visit here with Howell in 1933, several entire specimens of an Upper Cambrian *Agnostus* and a *Lingulella* were collected, and persistent searching will undoubtedly uncover other fossils.

Less than 1.5 miles to the north, and just east of the road to Georgia Plains, a cliff of crumpled Hungerford slate shows four bioherms lying close together, and ranging in width from 5 to 40 feet. They are more crystalline than is usual; Howell collected minute trilobites from one of them.

Georgia Center and west.—The village of Georgia Center, 5.5 miles southwest of St. Albans, on the highway to Burlington, is situated on the Georgia slate. The following section (Fig. 1, traverse IV) is along the line of the historic sequence known as the "Georgia Group", already described.

Leaving the north-south highway, one traverses the Georgia slates westward for about half a mile to a fork in the road, where the land rises steeply in an east-plunging limb, exposing the Rockledge breccia.

The writer first saw this fine exposure of the Rockledge breccia with Keith in August 1921, visited it the next year with Keith, Sayles, and Swinnerton, and has studied it several times since, as it gives the best record of the relation between the breccia and the Hungerford formation, on which it lies unconformably (Pl. 1).

On the northern branch of the road, the eastern limb of the Rockledge breccia rises in a steep arch, and thence, for about 1,000 feet westward, it lies in a flat undulating anticline, trending northward, with at least two undulations; between the undulations are lower places, which have been eroded through—windows, as it were—exposing beneath, first, a thin zone of the Hungerford slate, with dips as high as 15 to 30 degrees either east or west, then a lower dolomite about 10 feet thick, then the main mass of the Hungerford formation, which continues west for nearly 1.5 miles from the first outcrops of the Rockledge.

The decidedly cleaved breccia is here composed, in part, of pieces of the limestone lentils of all sizes up to 18 feet across; these pieces probably all came from the Hungerford slate, whereas three-quarters of the breccia consists of thin-bedded,

dark-blue limestone like that of the Gorge formation. Blocks, up to 3 feet in length, of the immediately subjacent Hungerford dolomites, are rare, and blocks of the slate itself are even rarer. The thickness of the breccia is about 30 feet, and it lies directly on the upper Hungerford slate, the intermediate dolomite zone, or the older Hungerford slate. These facts show clearly that the deposition of the Hungerford formation was followed by a long erosion interval.

On the southern branch of the road—the one leading to Georgia Plains—the Rockledge breccia is equally well exposed. At the west may be seen Hungerford banded slate; farther east, beneath the breccia, is a gray, massive, decidedly cleaved dolomite, which weathers a rusty red; still farther east, more Hungerford slate is present above the dolomite. All these beds are unconformably overlain by the breccia, which contains large blocks of the blue-white lentil limestone, in pieces up to 12 by 6 feet.

Longwell regards this as an undoubted unconformity between the Rockledge and the Hungerford formations. He described it in a letter to the writer, as follows:

"[The Rockledge breccia] lies nearly flat, or gently rolling, over a large area, and the Hungerford beds beneath are in a series of sharp folds, which were eroded somewhat before the Rockledge breccia was deposited. A bed of dolomite in the Hungerford serves as a good horizon marker and makes the fold structure clear. At some points the angular discordance between the Rockledge and Hungerford is 45° or more."

In the Rockledge breccia, Walcott says he collected *Huenella billingsi*, an Upper Cambrian brachiopod, and *Lingulella franklinensis* and *Ptychoparella adamsi*, both Lower Cambrian species (locality 28a).[87] The writer also thought he saw Lower Cambrian trilobites.

Keith does not agree with this chronogenetic interpretation. He said, in 1923, that the Georgia-Milton contact is also exposed at the locality to the northwest of Georgia Center, and that here the intervening Highgate [=Hungerford] slate appears to be cut out in places by erosion. The "Milton" dolomite, which he then regarded as of Lower Cambrian age, is "thrust up on a fault and locally capped" by the Rockledge breccia.[88] In 1932, he concluded that it is the Mill River, and not the Rockledge breccia, that is seen "resting on the Milton dolomite," here again regarded as of Lower Cambrian age.[89] Although the writer cannot prove, on the basis of fossils collected at this place, that Keith's correlations are wrong, the stratigraphic succession and the structure are unmistakably as follows: The Rockledge breccia here underlies the Georgia slate and overlies the Hungerford slate, the latter being on both sides of an interbedded dolomite.

The Hungerford slate is seen beneath, and to the west of, the Rockledge breccia for nearly 1.5 miles; this is the longest and best-exposed sequence of this formation. In this distance, at least three low anticlines were noted in the Hungerford, with the dip variable up to 30 degrees either to the east or to the west. If one takes half this distance at only 10-degree dip, the Hungerford slate has a thickness of at least 700 feet. In all this distance, no fossils have as yet been collected from the Hungerford formation, except near the base and in limestone lentils near the center of the slate series, but these are of Upper Cambrian age, and, as the slate is overlain

[87] C. D. Walcott: *Cambrian Brachiopoda*, U. S. Geol. Surv., Mon. 51 (1912) p. 189.

[88] Arthur Keith: *Cambrian succession of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 123.

[89] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 375.

by the Rockledge breccia, which has many more Upper Cambrian fossils, there can be no doubt of the identification.

The base of the Hungerford slate, here a dolomite breccia, lies upon the topmost dolomite of the Lower Cambrian, and not, as Schuchert[90] stated in his earlier paper, upon 57 feet of Middle Cambrian slate. This was pointed out to the writer by Howell in the field in August 1933. The contact may be seen near the point where the road that runs east to Georgia Center crosses another road leading north to the Parker farm (Fig. 1, loc. XXIX). The top of the Lower Cambrian in this exposure consists of a rusty-weathering dolomite, 30 to 40 feet thick (as on Parkers Cobble to the north), dipping about 15° E.; if projected across the north-south road, these strata would unite farther east with what the writer in 1933 called the Rugg Brook conglomerate. A re-examination of the latter exposure, 12 to 15 feet thick and dipping to the east, shows that the rock here is the same as the dolomite on the eastern slope of the hill to the west, and that it is overlain by a thin veneer of dolomite conglomerate at the base of the Hungerford. This dolomite conglomerate is derived from the Parker dolomite beneath. A horizon about 30 feet above the basal conglomerate has yielded a slender species of *Hyolithes*, fragmentary and entire specimens of two species of trilobites, and the hornlike fossil (*Bovicornellum vermontense*), which has proved to be the guide to the lower Hungerford slate. Raymond in his paper will discuss these two trilobites.

Fifty to sixty feet above the base of the Hungerford slate is a 10-foot zone with light-blue bioherms, forming the top of the little hill. There are at least ten of these bioherms, one 120 feet long, another 20 by 30 feet, and the rest smaller. This zone dips about 10° E.

This traverse, then, clears up what Walcott described in 1886 as the "Georgia Group". His divisions 1 to 6 are as he described them, and are of Lower Cambrian age. His division 7, said to have 3,500 feet of "argillaceous shales", embraces what is now called the Hungerford slate, with a thickness of at least 700 feet. Walcott's division 9, a "great lenticular mass," is the Rockledge breccia, about 30 feet thick. The great differences in thickness are due to Walcott's having mistaken cleavage for bedding. For more detail, see the Walcott section, given in the introductory portion of this paper.

Four miles southwest of St. Albans.—Nearly 4 miles southwest of St. Albans (Fig. 1, loc. XII), there is another east-west farm road (western half now abandoned), the greater part of which, east of Mill River, is occupied by Hungerford slate. There are several bioherms; the one north of the lane is 140 feet east-west by 120 feet north-south, and appears to be not more than 12 feet thick. Its growth is in vertical columns, each from 6 to 10 inches across (rarely as much as 18 inches), and each growing saucer-shaped, with the interspaces filled with a yellowish-weathering magnesian limestone (Pl. 5, fig. 1). No micro-structures can be seen, but such columnar growth can be due only to blue-green algae. South of the road and west of the barn lie three other smaller bioherms, now being uncovered through weathering. The largest of these lentils is 30 feet across; the other two are considerably smaller. From a pocket in one of these, numerous pieces of *Hemirhodon schucherti* and *H. viator* were collected. The slates have a few pieces of the bioherms, showing that the latter protruded above the general sea bottom and that the waves broke off pieces from time to time. These lentils also show scattered large grains of quartz sand. The largest lentil, more than 200 feet across and 20

[90] Charles Schuchert: *Cambrian and Ordovician stratigraphy of northwestern Vermont*, Am. Jour. Sci., 5th ser., vol. 25 (1933) p. 365.

feet thick, with trees growing on it, is farther south, in the woods. Farther east, to the north of the same lane, is a white sandstone lentil about 50 feet long on the strike of the formation, possibly 15 feet across the other way, and 5 feet thick. The dark slates show many small lentils of the same white sand rock, which have lengths up to 5 feet and thicknesses of 10 inches. They appear to be the result of current action, which brought together the wind-blown sand and piled it up as lentils.

Among these Hungerford lentils was found a glacially drifted block of the Rockledge breccia, which had in its various pebbles the following trilobites, described by Raymond: *Calvinella obsoleta, Hysteropleura macgerriglei, Onchonotus eminens, Distazeris acuta, Ucebia lata, Hemirhodon schucherti, H. minor, Stenocombus princeps, Lecanopleura interrupta, L. inflecta,* and *Tsinania scrinium.*

Less than three miles south of St. Albans.—Along the farm road west of St. Albans Hill (Fig. 1, traverse XXV), one sees the following sequence of formations: Less than a mile west of the main highway, there is exposed 10 feet of Mill River limestone breccia, which is composed of the usual blue-white bioherm limestone. The underlying Middle Cambrian St. Albans slate, about 70 feet thick, with 10 feet of Rugg Brook dolomite conglomerate at the base, is exposed farther west, as already discussed. The Hungerford formation, about 320 feet thick, crops out at intervals from the Mill River exposure, at the west, to that of the Rockledge breccia about 1,000 feet west of the main highway; these mudstones are the usual much-banded and highly cleaved dark slates, and they dip from 8° to 12° E. The section terminates eastward with the Georgia formation.

Mayre Ledge, two miles south of St. Albans.—The Rockledge breccia is well shown in a large exposure 2 miles south of St. Albans (Fig. 1, loc. XXX) and appears to have a thickness of 30 feet. The cleavage, with a dip steeply eastward and the boulders drawn out, gives a false impression of bedding, but the latter is clearly shown to be nearly horizontal in the western cliff. Blocks of the Hungerford bioherms measure as much as 20 feet across. Some dolomite masses are present. Most abundant are slabby limestones like those of the Gorge formation. The cement of the conglomerate is the usual coarse round-grain sand, which is present in thick beds at the southwest.

Fossils noted here were rare Cryptozoons and trilobites, of which Raymond identifies *Maryvillia triangularis.*

Less than a quarter of a mile southeast of Mayre ledge, a series of Rockledge breccia outcrops extends south for five-eighths of a mile to the road east to St. Albans Hill. These beds also include many large blocks of the marbleized bioherms out of the Hungerford slate, some of them 60 feet in length. One place in this conglomerate shows a local filling of the usual coarse-grained white sandstone, 30 feet thick.

Adams pasture, St. Albans.—This is the type area (Fig. 1, traverse III) for the Middle Cambrian (St. Albans slate) already described in detail. The slate is overlain disconformably by the basal Upper Cambrian Mill River limestone breccia, here about 15 feet thick. The breccia is composed mainly of angular pieces of bluish white limestone lentils, of all sizes up to 20 feet long; few, if any, of the slabby dark-blue pieces like those of the Gorge formation are to be seen, although rare pieces of a dolomite, some of which are 8 inches long, occur; according to the evidence of the fossils, there are also pieces out of the St. Albans and the Parker formations. At the base of the Mill River breccia are bedded sandstones, up to 18 inches thick, composed of the usual rounded sand grains and dipping 9° E.

Above the breccia is the regulation banded Hungerford slate, the thickness of which is not determinable at this point. However, east of the city of St. Albans the overthrust Lower Cambrian of the Eastern Sequence is seen and Marcou notes the finding of *Olenellus* within the village. If, therefore, the Hungerford slate continues up to the main north-south street of St. Albans, the thickness of the formation in this area exceeds 600 feet.

Three and a half miles north of St. Albans.—A large limestone lentil here (Fig. 1, loc. XIV) is surrounded by Rockledge breccia; it is 65 feet long, 35 feet wide, and 15 feet thick. Another block, 20 feet long, lies to the north of it. The breccia dips 10° E., striking north, and surrounds the lentils. On the west, the breccia pile is jointed off, showing that the lentil rests on thin-bedded, sandy, slightly conglomeratic (with small slate and limestone pieces), and fossiliferous dolomite. These basal beds are 10 feet thick or more, are of the Rockledge *in situ*, and have yielded ten species of trilobites listed in Table 2.

A short distance to the west of the conglomerate pile, black slates, which are also conglomeratic, are exposed, and these appear to be the top beds of the Hungerford formation, as will be demonstrated by a locality farther north, the next to be described. Still farther west, undoubted Hungerford slate rests upon the Middle Cambrian (St. Albans).

A close examination of the weathered surface of this large marbleized bioherm shows flow structure, with small pieces of reddish dolomite moved about in the limestone. When did this metamorphism take place? Probably some of it is original, due to reef growth and diagenetic action, but most of the change must have taken place during the Taconian orogeny, when the entire region was folded and overthrust. In fact, at all the places where Rockledge breccia is exposed, there is more or less cleavage developed, and at several places the limestone pieces are drawn out. The time of making of these internal structures, and their causation, were discussed with many geologists while they stood on the bioherm, and it was agreed that the bioherm may be in the place of its origin, or it may have been moved by gravity a short distance; probably, during its growth, the lentil was more or less altered diagenetically and, in places, dolomitized, and it finally was fully metamorphosed and cleaved during the orogeny that affected the whole region.

Some hundreds of feet south of the lentil, and along the western cliff face of the breccia, a fine section reveals the nature of the basal *in situ* part of the Rockledge breccia, here mainly a sandstone composed of rounded sand showing bedding and current action. In this sandstone bits of trilobites are found, and to the north, beneath the lentil, are the equivalent sandy dolomites with trilobites, showing that the Rockledge is a marine deposit *in situ*. The following sequence was made out (Pl. 6, fig. 1):

Bed	Thickness Feet	Inches
4. Limestone breccia, squeezed and cleaved, about	7	
3. Sandstone of round grains, with a few limestone pieces. Sandy dolomite pebbles are also present	1	2
2. Sandstone, with many small limestone pebbles	2	
1. Limestone breccia, with most of the pebbles dipping southwest, showing foresetting	3	
Total	13	2

Below are black slates thought to be of Hungerford time.

In the overlying Georgia slates, small limestone pebbles are found scattered through a thickness of 50 feet. Pieces of the lentils, up to 2 feet across, are exposed rarely, and one block was noted 10 feet long. This indicates that deposition was continuous from Rockledge to Georgia time.

Rockledge.—From afar, one sees gleaming in the sun the large white limestone bioherm (Pl. 5, fig. 2) on the estate known as Rockledge, the home of Doctor Janisen, about 4 miles north of St. Albans (Fig. 1, loc. XIII). It appears to be at least 170 feet long by 110 feet wide and more than 10 feet thick. There is another block, to the south of the larger mass, which is 90 by 50 by 10 feet and may be a part of the same lentil. The lentil lies in place, upon banded black slate of the Hungerford formation (contact seen to the west), and is surrounded by the regulation Rockledge breccia, exposures of which are seen east to the doctor's home. At first, the idea was entertained that the lentil might be a mass moved out of the Lower Cambrian Parker slate, and only after other bioherms had been found in the Hungerford slate to the north, and especially to the south, did Keith and the writer come to the conclusion that it is in place. The Rockledge estate and the terrane for a mile to the south form the type area for the Rockledge breccia.

This lentil shows the same make-up and flow structure as the one less than a mile south, with thin irregular pieces of yellowish dolomite moved around in the limestone.

The Rockledge lentil protruded above the surface when the sea invaded the area and deposited the Rockledge breccia. The sea waves pounded upon the bioherm and broke away parts of it, piling these fragments, and a great mass of thin-bedded, dark-blue limestones derived from the Gorge formation, about, and over, the great lentil, to make the Rockledge breccia.

Fossils were collected at the east base of the Rockledge lentil in the sandy paste of the Rockledge breccia. These are of the trilobite *Cephalocoelia ovoides* Raymond. The sandstone was about 3 inches thick and 4 feet long. Elsewhere, Prindle, Raymond, and the writer collected trilobites in the slabby sandy limestones of the Rockledge breccia; these proved to be: *Acrohybus argutus, Stenelymus kobayashii, Greylockia minuta, Kaninia ? platys, Maryvillia triangularis, Homodictya imitatrix, Protillaenus marginatus, Coleopachys strix,* and *Bynumia leptogaster.*

Pieces of the bioherm limestone are scattered in the basal 50 feet of the Georgia slates.

A short distance to the south of Rockledge is the cross-road west to Fonda quarry (Fig. 1, traverse XXII), and to the south of the barns at the intersection of the highways may be seen a cliff exposing, at the top, the Rockledge breccia, beneath which are shales, which appear to be of the Georgia formation. These shales, in turn, lie unconformably on the older Hungerford slate. The detail here is as follows: At the base of the cliff is the regulation blue-black banded Hungerford slate, with an irregular upper surface—the unconformable line of contact. Above this contact the Georgia formation begins with gray slates, without a trace of limestone inclusions in the first 5 feet. In the next 5 feet, small scattered limestone pieces are more common, but none of them exceeds a foot across, and all appear to be pieces of broken lentils. In the next 10 feet the inclusions are still more common, and then the regulation Rockledge breccia is developed.

To the west of Rockledge, the Hungerford slate has a thickness of 215 feet, and beneath it lies 8 feet of the Mill River breccia, separating it from the Middle Cambrian (St. Albans) slate.

Four and a half miles north of St. Albans.—The sequence 4.5 miles north of St. Albans (Fig. 1, traverse VIII) is exposed along another cross-road, which connects the Highgate and the Swanton roads, a short distance to the north of Rockledge. The sequence begins 1,600 feet northwest of the Highgate road, where the top of the Parker slate crops out, as already described. Overlying the Parker slate is a sandy, rusty-weathering dolomite, several feet thick, which probably is the terminating bed of the Lower Cambrian at this locality. The Middle Cambrian is absent, although it is exposed half a mile south, to the west of Rockledge. Unconformably upon the dolomite rests 6 feet of dolomite breccia, composed of slabby and massive pieces probably derived from the Lower Cambrian; it holds the stratigraphic position of the Mill River breccia, but is entirely unlike the latter in appearance.

Above the breccia lies from 10 to 15 feet of Hungerford slate, and above this, in turn, is a sandy dolomite breccia, from 3 to 5 feet thick, with blocks up to 18 inches across. Between these breccias lies a local lens of white quartzite, 2 feet thick. *Bovicornellum* is common in slate beds above the second conglomerate, and again 50 feet higher.

The distance from the basal breccia eastward to the top of the Hungerford is 1,600 feet, and, at an average dip of 10° E., the thickness of this formation would be about 275 feet, or considerably less than that usually present.

The Rockledge breccia is exposed along the Highgate road and overlies the Hungerford slate in unconformable relation.

Six miles north of St. Albans.—The traverse 6 miles north of St. Albans begins at Skeels Corners and goes west for about a quarter of a mile. A long exposure of the Rockledge breccia, about 30 feet thick and dipping 25° E., which in every way looks like this breccia as it is exposed a short distance to the south, is to be seen north of the Corners and back of the two houses. It is underlain unconformably by a local variation of the Hungerford formation—a gray, massive, sandy dolomite, much veined with calcite and quartz, pieces of which are also found in the breccia. The dolomite likewise has, scattered through it, small angular pieces of a finer-grained dolomite, some of them 10 inches long. These dolomite strata, 20 to 25 feet thick, appear to be a local recurrence in the upper part of the Hungerford formation, apparently holding the same position as the dolomite beneath the Rockledge breccia west of Georgia Center.

These correlations are not acceptable to Keith, who says that the difficulty in separating what the writer here calls Rockledge breccia from the Mill River breccia "is most considerable for a few miles from Highgate Falls south to Skeels Corners, for the Mill River conglomerate together with a part of the Milton dolomite [the writer's 25-foot dolomite bed, which he thinks is of Lower Cambrian age] is repeated by a thrust fault and now lies on top of the Highgate [=Hungerford] slate. The thrust fault is well exposed in the gorge at Highgate Falls," but this thrust breccia is at the base of the Highgate, whereas the Mill River one is at the base of the Hungerford slate.[m]

Neither Longwell nor the writer could see the thrust faults beneath the Rockledge breccia at Skeels Corners. The Rockledge north of Skeels Corners rests on a dolomite of the Hungerford, as it does west of Georgia Center.

Just beneath the dolomite are almost cross-bedded, dark slates, which weather rusty and dip 10° E. Considerably farther west, in banded slates, Prindle found *Bovicornellum vermontense* of the Hungerford slate. No Middle Cambrian slates

[m] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 375.

are known this far north; the last outcrop proved by fossils is 2 miles south and shows a thickness of only 60 feet. Farther west, the top of the Lower Cambrian Parker formation shows about 35 feet of hard strata, with sandy dolomites in the upper portion, underlain by white sandstone which has a basal dolomite breccia about 5 feet thick. Still farther west, the regulation Parker slates are exposed.

Six and a half miles north of St. Albans.—The traverse now to be described starts farther north, at Highgate Falls, goes 2 miles south, and thence west (Fig. 1, traverse IX) for a mile, along a lane leading to the Donaldson farm.

South of Highgate Falls, nothing is exposed, except Champlain sands, for more than half a mile, and then there is a rise in the land of about 40 feet, but no sig-

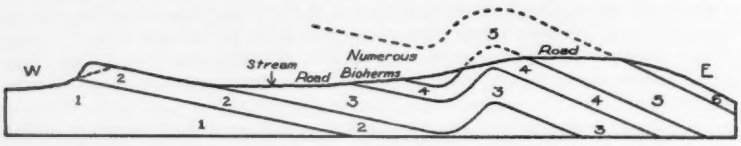


FIGURE 12.—*Sequence from Donaldson farm east to Highgate road*

Distance, 1 mile (Fig. 1, traverse IX). (1) Parker slate overlain by (2) thrust breccia at base of Hungerford slate. (3) Lower Hungerford with horn-like fossil; (4) local dolomite in the Hungerford; (5) upper Hungerford; (6) Rockledge breccia.

nificant rock exposures appear until one approaches Hungerford Brook, a mile south of Highgate. This is the type area for the Hungerford formation (Fig. 1, loc. XXVIII), and gives a fine display of that highly cleaved dark-blue slate. A few trilobites, as yet undescribed, were found here many years ago by Billings, and many more have been collected in recent years by Howell, who intends to describe them. The slate in the brook to the east of the Highgate road dips east, but, on the falls to the west of the road, the dip is steeply to the west. This marked anticline can be traced from the brook south for at least a mile, to the Donaldson cross-road, along which the whole thickness of the Hungerford formation is more or less well exposed.

Fine exposures of the Rockledge breccia, of the usual composition and thickness (30 feet), are to be seen between the first two houses on the Highgate road north of the cross-road (Fig. 12). To the west of this breccia, for about 850 feet, the only strata visible are highly cleaved Hungerford slate in the anticline already mentioned, with dips steepening to 20° E. and to 60° W. Between the Highgate road and the west-facing cliff of this anticline lies a bioherm, 50 feet across, and, east of it, a bed of sandy dolomite, 10 feet thick. The thickness of the upper Hungerford slate exposed here is estimated at 125 feet.

Going west from the Highgate road, along the Donaldson lane for 750 feet, one comes to the anticline in the Hungerford slate already described (Fig. 12). The west-facing ridge beneath the slate exposes a sandy hard dolomite—almost a sandstone—about 15 feet thick, lying upon a fine-grained, hard, sandy, dolomitic slate, approximately 20 feet thick, dipping steeply to the west. In the next 400 feet, the road crosses a small syncline showing the Hungerford slate just mentioned, and then the same sandy dolomite seen to the east reappears, with a thickness of 35 feet and a dip of 20° E. This dolomite is, in places, an intraformational dolomite breccia. Above it, in the slate, an *Agnostus* was collected, and the same form is found elsewhere in the Hungerford. A great thickness of Hungerford slate with the same dip lies beneath the dolomite, and *Bovicornellum* was found in it.

Farther west and to the north of the road, in the woods, the lower Hungerford slate has many bioherms, of all lengths up to 300 feet, which cover an acre or two. Continuing west for a quarter of a mile, the road crosses a small stream and swamp on Champlain clay, and then, in harmony with the easterly dip slope of the Lower Cambrian Parker slate, the land rises all the way west to the next north-south road, on the land of Hugh Donaldson.

Striking outcrops of the basal Hungerford thrust breccia, about 15 feet thick, overlie the easterly dipping Parker slate. The breccia has large blocks of the white lentil limestones, as much as 4 feet across, and others of Parker slate, 3 feet long. In places, most of the blocks are of sandy dolomite, and two that measure 15 by 10 by 10 feet were seen; these weather salmon color and appear to have been derived from the Parker formation. In this breccia, down the eastern dip slope, lies a bluish white lentil, now broken up by weathering, which was originally 100 by 50 feet; it looks as if it were resting on the Lower Cambrian. This is the most extensive exposure of a thrust breccia known in northwestern Vermont, but, curiously, no marked metamorphism, crushing, or slickensiding was seen.

The traverse across the Donaldson road, from east to west, reveals the following formations:

	Approximate Thickness Feet
UPPER CAMBRIAN	
Rockledge breccia	30
Break and disconformity.	
Upper Hungerford slate, with a bioherm and a 10-foot bed of dolomite	125
Intermediate sandy dolomite and dolomitic slate	35
Lower Hungerford slate, with many bioherms	410
Hungerford basal thrust breccia	30
Total for the Hungerford	600
Thrust unconformity	
LOWER CAMBRIAN PARKER SLATE OF DONALDSON FARM	

Highgate Gorge.—The Highgate road crosses the Missisquoi River at Highgate, and on the higher north side of the stream is the village of Highgate Falls. Below the falls and dam the Highgate formation is well exposed (see Ordovician section at Highgate Falls).

Below the Highgate formation the descending section of the Gorge formation is as follows:

UPPER CAMBRIAN, UPPER GORGE limestones and dolomites (145 feet).

Bed		Thickness Feet
25. Dark-blue limestones and limy shales in beds one-fourth to three-fourths inch thick. These strata were deformed when the breccia bed slid across them, so that the zone is irregular in thickness. Into these beds have been shoved several blocks of dolomite, 1-2 feet long, indenting the beds below		2 – 4
24. Massive conglomeratic dolomite	Irregularly bedded strata with an average thickness of 10 feet	0 – 4
23. Wedge of blue thin-bedded limestone		0 – 4
22. Massive dolomite conglomerate		0 – 2
21. Blue thin-bedded limestone		1 – 5

Bed		Thickness (Feet)
20.	Massive conglomeratic dolomite	1.5– 4
19.	Thin-bedded dolomite, dolomite, and slate. These strata are partly hidden by a rock dump. East of the dump, this zone, about 6 feet thick, has a lens of intraformational breccia, which increases to 3 feet in thickness; near the top, a few trilobites have been collected (*zone 3*). West of the dump the same zone reappears, with a thickness of 10 feet, and furnishes the best collecting ground for many kinds of trilobites and a few brachiopods (*zone 2*)	6 –10
18.	Intraformational dolomite conglomerate (18 inches). Thin-bedded limestone and dolomite (18 inches). Intraformational breccia (2 feet). Thin-bedded limestone and dolomite (5 feet). To water's edge	10
17.	Downstream, gray dolomite in two or three beds. The great flood of 1927 cleaned out all the loose material, and every foot of the lower section could be seen, and most of the strata are still visible. The beds are irregular, all being shallow-water deposits with short lenses, especially of sandstones, deposited by current action.	2
16.	Black dolomitic slate and dolomite (up to 3 feet), all in lens deposits. Near the base, rare *Lingulella* and *Acrotreta*	20
15.	Banded muddy sandstone (2 feet), sandstone (15 inches), massive, irregularly bedded dolomite, with black flat dolomite pebbles embedded in it (4 feet)	8
14.	Black thin-bedded dolomitic slate, with local lenses of gray dolomite	8
13.	Massive, current-rippled sandstone	6
12.	Black dolomitic slate, with thin and thick lenses of gray dolomite	18
11.	Massive sandstone lens	6
10.	Black dolomitic slate, with a limestone bed, 1 inch thick, replete with trilobites (*zone 1*), and also yielding two small species of *Lingulella*	8

Keith[88] regards all the beds down to this horizon as belonging to the Gorge formation. The dolomites below he correlates with the Milton, which he refers to the Lower Cambrian. He considers the two parts to be separated by a thrust fault. On July 18, 1932, Longwell, Dunbar, and the writer re-examined the Highgate Falls section, in an endeavor to find this thrust fault, but without success; the writer, therefore, believes that the entire sequence throughout the Highgate gorge is unbroken, except at the thrust breccia. Accordingly, it is his belief that the upper Gorge continues downward as follows:

Bed		Thickness Feet
9.	Massive sandstone lens	1.5
8.	Black dolomitic slate, with sandstone layers	4
7.	Sandstone	1
6.	Black dolomitic slate, interbedded with lenses of sandstone	7.5
5.	Intraformational dolomite conglomerate, with flat black pebbles	3

[88] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 375.

Bed		Thickness (Feet)
4.	Blue and black dolomitic slate	4
3.	Gray, massive, intraformational dolomite breccia, with flat black pebbles	12
2.	Pocket of black dolomitic slate, containing *Lingulella*	5

Keith[98] thinks there is a "fair prospect" that Middle Cambrian strata can be identified in one of the slate beds. The writer, on the contrary, is convinced that there is no such possibility in the Highgate gorge, especially because no Middle Cambrian is known to the north, and the first known outcrop of this formation, with fossils, is 4.5 miles to the south, west of Rockledge.

The upper Gorge division (Pl. 2) is exceedingly irregular in its deposits, for at many horizons the strata swell quickly from thin to thick beds, or change into intraformational conglomerates or from sandstone into black thin-bedded dolomites. In addition, it includes many intraformational dolomite breccias with thin flat pieces of limestone or dolomite. The sequence as a whole gives the impression of being the result of localized deposition on a sea bottom where local slips were intermittent, and where current action was strong. The sandstones, which are the most irregular in thickness and development, appear to be local current accumulations among the dolomite growths, and the thickest beds vanish in a distance not exceeding several hundred feet. The black thin-bedded dolomites are more persistent, and must have formed locally in depressions on the sea bottom, throughout the entire time of deposition of the Gorge formation. They were then in part broken up into angular pieces, possibly by earthquakes but more likely by shallow-water current waves, and entombed in all the deeper-water dolomites, as intraformational breccias

Downward, the section is:

LOWER GORGE (75 feet) massive sandy dolomite, replete with black flat-pebble inclusions, up to a foot in length. In places, the dolomite is so sandy as to be almost a sandstone.

Bed		Thickness Feet
1.	Basal intraformational conglomerate, making a massive pebble dolomite. It is a tangled brecciated mass, bound together by a muddy blue-black dolomitic paste. Four feet is exposed above the river and more is beneath. This same basal conglomerate is also seen about a mile north, where it lies on green and black slate ("Hungerford") with *Dictyonema*	6+

Local convulsions of the sea bottom may have taken place during all the time when the Gorge formation was being deposited, as seems to be indicated by the endless repetition of intraformational conglomerate beds and interbedded sandstones. Possibly, the area was being faulted, resulting in submarine cliffs, and intermittent uplifts and earthquakes caused beds to slide on the sea bottom. In this way, and with the further help of tidal waves (tsunamis), the numerous intraformational conglomerates of the Gorge formation may have been made. In any event, intraformational conglomerates were being formed, especially throughout the time of Gorge deposition, but not during Highgate time. The difficulty in this assumption, however, is that faulting is nowhere visible in the immediate area.

[98] *Op. cit.*, p. 374.

The A1 excursion of the Sixteenth International Geological Congress of 1933 studied the Highgate Gorge sequence, and it was the opinion of O. T. Jones, of Cambridge University, who is familiar with similar phenomena in the Sutherland area of Scotland, that the various breccia beds could be explained as due to faulting or at least to earthquakes. The Sutherland area is described by Bailey and Weir.[94]

ORDOVICIAN

FORMATIONS AND FAUNAS

General Statement.—In the Central Sequence, northwestern Vermont has the following formations of Ordovician age: (1) Highgate formation, (2) Corliss limestone breccia, and (3) Grandge slate.

Highgate Formation.—The type area of the Highgate[95] formation (Fig. 1, locs. XI and XXVII) is the upper sequence in the gorge of the Missisquoi River at Highgate Falls, from which point these strata extend a short distance north into Canada. In the type area the formation begins with a thick limestone breccia, made by a tectonic thrust movement from east to west (Pl. 6, fig. 2). This breccia has heretofore been wrongly correlated with the Mill River breccia, which lies locally at the base of the Hungerford slate of the *early* Upper Cambrian, whereas the thrust breccia and the younger Highgate formation overlie the Gorge formation, which is of the *late* Upper Cambrian.

Above the thrust breccia lies a series of alternating impure thin-bedded limestones, rare dolomites, and shales (= lower Highgate), and these merge upward into well-banded slates (= upper Highgate). In places, the banded slates are absent, apparently as a consequence of later erosion. The formation covers a great part of Highgate township and is generally strongly cleaved.

The lower Highgate (Pl. 7) consists of interbedded thin layers (from an inch up to several inches in thickness) of blue limestones, rare gray dolomitic beds, and slates, giving the formation a well-bedded character and, where folded, a picturesque appearance. The higher slates are also banded by paper-thin siltstones. At Highgate Falls, a massive dolomite, very much like the one at the base of the Gorge formation farther downstream, is overthrust upon the top of the Highgate slates. Accordingly, the Highgate formation at the falls has a thickness of about 300 feet.

The banded nature of the upper Highgate slate was best seen in 1922 at a point about 3 miles northeast of Highgate Center (Fig. 1, loc. XXXI), where the highway up the little hill from South Gore school was being widened. The cleavage here is at right angles to the bedding of the blue-

[94] E. B. Bailey and J. Weir: *Submarine faulting in Kimmeridgian times: East Sutherland,* Roy. Soc. Edinburgh, Tr., vol. 57, pt. 2 (1932) p. 429-467.

[95] Arthur Keith: *Cambrian succession of northwestern Vermont,* Am. Jour. Sci., 5th ser., vol. 5 (1923) p. 114-115.

FIGURE 1. LOWER HIGHGATE LIMESTONE
Thin-bedded and deformed limestones, dipping downhill. Grandge farm, near Canadian boundary (Fig. 1, loc. XV). Photograph by G. Marshall Kay.

FIGURE 2. LOWER HIGHGATE LIMESTONE
Close view of figure 1. Note thin veneer of Grandge slate between hammers. Photograph by G. Marshall Kay.

FIGURE 1. CONTACT BETWEEN GRANDGE AND HIGHGATE FORMATIONS
Angular unconformity (produced by Quebec orogeny) between basal limy and fossiliferous mudstones of Grandge formation and banded limestones of lower Highgate; contact at hammer head. Grandge farm (Fig. 1, loc. XV). Photograph by G. Marshall Kay.

FIGURE 2. CONTACT BETWEEN GRANDGE AND HIGHGATE FORMATIONS
The same contact, farther north. Photograph by G. Marshall Kay.

black slate, which plainly showed the extraordinary banding, consisting of thin yellowish white laminae of fine silt or clay. As many as seven of the lighter-colored bands were counted in a space of three-eighths of an inch. A specimen of this slate was given to W. H. Bradley, of the United States Geological Survey, who sent the following report to the writer on June 11, 1932:

"I have found this slate to be laminated and that the lamination is in part at least an original sedimentary structure, but how to interpret it is quite another matter. . . . Some of the laminae consist of silt and some of clay which has been metamorphosed, and some consist of silty clay which apparently contains organic matter. The laminae differ considerably in thickness, and because there seems to be no systematic arrangement of them or systematic alternation of lithologic types, I do not see how they can be related to any particular natural cycle like the seasons. . . . The type of lamination shown by the Highgate slate seems to me impossible to distinguish from a lamination that might be produced by storms or currents or some other changes in the manner of [marine] sedimentation whose time value cannot be determined."

At various places near Highgate Center, but more especially near the Canadian boundary (Grandge farm, Fig. 1, loc. XV), the writer and others have collected trilobites, which have been described by Raymond.[96] The 37 species come from the lower Highgate, and are listed in the first two columns of the table on page 1072.

Of the 37 species in the lower Highgate, only 2 are found in the Gorge formation below—namely, *Lloydia seelyi* and *Maryvillia triangularis*. Three species continue into the younger Grandge strata—namely, *Lloydia saffordi, Triarthroides cyclas,* and *Leiostegium obtectum*. These 37 species are grouped in 26 genera, and of these genera no fewer than 16 are known elsewhere in the Ordovician. The following, according to Raymond, are more especially Ordovician types: *Triarthroides, Shumardia, Warburgella, Metoptogyrus, Metapliomerops, Leiostrototropis, Plumulites,* and the four species of *Isochilina*. Significant also are the Cheiruridae, forms "characteristic of the oldest Ordovician deposits in Scandinavia and Manchuria."

The "surprising and significant feature" of the foregoing table is, Raymond says, the presence of three forms that were previously known in the Shumardia limestone at Lévis, Quebec. These are *Plumulites mobergi, Shumardia granulosa,* and *Warburgella angelini.*

"Since it is in this zone at Lévis that the first of the diplograptid graptolites appear, it has been generally considered to be very high in the Beekmantown, if not the oldest of the Mid-Ordovician series. As shown by the above analysis, the preponderance of evidence is that this is an Ordovician rather than a Cambrian fauna."[97]

[96] P. E. Raymond: *New Upper Cambrian and Lower Ordovician trilobites from Vermont,* Boston Soc. Nat. Hist., Pr., vol. 37 (1924) p. 389-466; reprinted, with a few changes, in Vt. State Geol., 14th Bien. Rept., 1923-1924 (1924) p. 137-203; *Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont,* Geol. Soc. Am., Bull., vol. 48 (1937). In press.

[97] P E. Raymond: *Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont,* Geol. Soc. Am., Bull., vol. 48 (1937). In press.

TABLE 3.—*Fossils (mainly trilobites) of the Highgate formation*

Fossil	Lower Highgate		Lévis, Quebec	Genera of Ordovician affinity
	Around Highgate Center	Grandge farm		
Lingulella acuminata (Conrad)	x			
Agnostus sp.	x			x
Hypagnostus sp.	x			
Leptopilus declivis Raymond	x			
Cholopilus vermontanus Raymond	x			x
Leiostegium puteatum Raymond	x	x		x
L. cingulosum Raymond	x			
Maryvillia triangularis Raymond	x			
Also in upper Gorge				
Lloydia seelyi (Walcott)	x			
L. saffordi (Billings)	x	x		x
Also in Grandge formation				
Pilekia apollo (Billings)	x			x
Hemigrapsis sp.	x			x
Metapliomerops extenuata (Raymond)	x			x
Lecanospira sp.		x		x
Dirachopia sp.		x		
Plumulites mobergi Clark		x	x	x
Isochilina puteata Raymond		x		x
I. tenuifilum Raymond		x		
I. eximia Raymond		x		
I. bilabiata Raymond		x		
Shumardia granulosa Billings		x	x	x
Entomaspis clarki Raymond		x		
Also in Upper Cambrian				
Triarthroides cyclas Raymond		x		x
Also in Grandge formation				
Diplozyga striata Raymond		x		
Warburgella plana Raymond		x		x
W. angelini (Billings)		x	x	x
Perischodory grandgei Raymond		x		
Lloydia brevis Raymond		x		
Leiostegium elongatum Raymond		x		
L. obtectum Raymond		x		
Also in Grandge formation				
Diplapatokephalus multispinosus Raymond		x		
Metoptogyrus grandgei Raymond		x		x
Strototropis laeviuscula Raymond		x		x
S. elevata Raymond		x		x
Leiostrototropis phlegeri Raymond		x		x
L. laevis Raymond		x		
Metapliomerops latidorsatus Raymond		x		x

Shumardia (*S. pusilla*) also occurs in the Deepkill shale of the Lower Ordovician of New York.

The writer now agrees with Raymond that the Highgate formation as restricted and re-defined in this paper should be referred to the Ordovician, correlating with the European Ceratopyge beds. The Highgate lies unconformably beneath the Grandge slates, another member of the

Lower Ordovician; the trilobites found in the basal stratum of the latter led Raymond to conclude in 1924 that the Grandge fauna also indicated a Lower Ordovician age.

For the position of the Ceratopyge beds in the geological time table, the reader is referred to a recent paper by Störmer.[98]

Post-Highgate Emergence and Orogeny.—An angular unconformity between the lower Highgate and the overlying Grandge slates was discovered by Keith in 1922 in the Rock River valley on the Grandge farm, 4.5 miles N 10° E of Highgate Center and 0.5 mile south of the Canadian boundary. The crustal movement thus indicated has since been called the Vermont orogeny by Schuchert and Dunbar,[99] but Kay [100] has recently re-named it the Quebec disturbance, to agree with the land known as the Quebec barrier. It is an early phase of the Taconian disturbance (Pls. 7 and 8).

The formations are here folded into a syncline, and immediately above the unconformity a calcareous zone in the Grandge slate contains many brachiopods and trilobites of Beekmantown kinds, discussed in connection with that formation, which appear either as actual tests or as molds in the weathered surfaces of the rock.

The species on either side of this angular unconformity are not markedly different, and both the Highgate and the Grandge faunas are of Lower Ordovician age. The lost interval is therefore not of great significance.

This unconformity was seen in other places on the same farm in the next quarter of a mile to the north, and it is especially well shown, and decidedly angular, on the eastern slope of the Highgate formation. The base of the Grandge formation cuts across different levels of the older, eroded Highgate, and all the upper Highgate is absent. In several places, the basal Grandge strata reveal small pockets of limestone conglomerate with pieces of Highgate limestone, but in none of these was the thickness greater than 1 foot, nor was the pocket more than 5 feet across; these limestone conglomerates are not to be correlated with the Corliss conglomerate. Higher in the Grandge slate it is not uncommon to see pieces of the older Highgate; some of them are 2 feet long, and one of them, near the east-west road, is 10 feet long. It is a puzzle to determine how these large blocks of the Highgate got into the muds of the Grandge formation, unless there were low cliffs nearby from which they were broken by sea waves or earthquakes.

In July 1932, during a visit to the Grandge farm, Longwell, Dunbar, and the writer studied this unconformity between the Highgate and the Grandge formations, and it was the general agreement that an unmis-

[98] Leif Störmer, *in* Olaf Holtedahl, *et al.: The geology of parts of southern Norway*, Geol. Assoc., Pr., vol. 45 (1934) p. 332.

[99] Charles Schuchert and C. O. Dunbar: *Textbook of historical geology* (1933) p. 124-126.

[100] G. M. Kay: *Stratigraphy of the Trenton group*, Geol. Soc. Am., Bull., vol. 48 (1937) p. 29.

takable angular and erosional unconformity is present. Furthermore, the dips and strikes in the two formations vary. According to Longwell's readings, the bedding in the Highgate limestones strikes north-south and dips 18° E., whereas the younger Grandge slate strikes N 35° W and dips 27° NE. At one point 200 yards south of the boundary, the angular discordance between the two formations is 30 degrees.

In September 1934, the writer re-visited the Grandge farm, this time with T. H. Clark and G. Marshall Kay, and the party discovered the most striking angular unconformity yet known between the Highgate and Grandge formations; this is shown in Plates 7 and 8, which are from photographs taken by Kay. At this place, 5 feet of Highgate is cut out in 12 feet across the unconformity.

The older Highgate is not much cleaved, but the Grandge shows decided cleavage, which dips 75° E. The Grandge is clearly an overlapping and transgressive deposit, with its beds fitting closely into all the irregularities of the Highgate strata below. The contact is not a thrust plane, for one sees no evidence of thrust phenomena. The evidence, on the contrary, shows with certainty that the Highgate was folded, and that all the upper Highgate slates were eroded away before the Grandge formation was deposited. Later, during the Taconian orogeny, both formations were folded together.

This same Quebec unconformity, exposed over a wider area, is again to be seen 1 mile to the north, in Canada, at the falls of Rock River, on Lot 123, where a country road crosses the stream. In 1865, Herrick's mill stood at the falls. The decidedly cleaved Grandge formation, here showing a thickness of about 20 feet, lies unconformably on the folded and slightly cleaved, thin-bedded lower Highgate limestone, of which about 6 feet is exposed. The bedding in the Highgate strikes N 10° E and dips 20° E, whereas that in the Grandge formation dips more steeply (about 45 degrees) east, according to Longwell.

Corliss Breccia.—About 5 miles N 10° E of St. Albans in a straight line (Fig. 1, loc. XXXII), a conspicuous limestone breccia, known locally as Corliss ledge, is to be seen, extending about 1500 feet north and south. This is the type locality for the Corliss breccia. No bedding can be made out because of the large size of the pieces and the marked cleavage, and the dip appears to be less than 20 degrees to the east. The matrix has far less sand than the Rockledge breccia. In general character, the Corliss breccia resembles the Rockledge breccia at the base of the Georgia series; its blocks range up to 25 feet across, but most of the smaller pieces appear to be out of the lower Highgate instead of the Gorge formation. The thickness may not exceed 30 feet. The breccia

is presumably underlain by the Highgate slate and overlain by another dark slate, which may be the Grandge formation.

Near the center of Corliss ledge, the writer collected, in a pebble, *Clarkella* n.sp., and another brachiopod near *Finkelnburgia*. Keith found here in 1921 several specimens of small and not well-preserved, closely septate cephalopods, similar to those occurring in the limestone near Brandon, which are of Lower Ordovician age. In 1923, Raymond and the writer collected more cephalopods from pebbles, but none of these is determinable. Of trilobites found here, Raymond describes the following:

Symphysurina globosa Raymond
S. parva Raymond
Symphysurinella corlissensis Raymond
Symphysuroides latus Raymond

The general aspect of this trilobite fauna, Raymond says, "is distinctly Ordovician, for most of the fragments can be identified as remnants of asaphid trilobites. . . . All the evidence to be derived from the trilobites indicates that the pebbles in the Corliss are of Ordovician age." [101]

Grandge Slate.—At a place 4 miles north of Highgate Center and about a quarter of a mile east of Rock River (Fig. 1, loc. XXXIII), in a calcareous sandstone 75 feet above the base of the Grandge slate, Keith and the writer collected a number of brachiopods, identified by G. Arthur Cooper as *Archaeorthis hippolyte*, *A.* cf. *A. electra*, and *Finkelnburgia* sp. They are of Lower Ordovician age. Originally the writer erroneously referred this horizon to the Middle Ordovician, but Raymond has shown, on the basis of the trilobites found farther north, that the age is Lower Ordovician.

Half a mile farther north, on the Oliver Grandge farm, and at the very base of the Grandge slate, in the type locality for the formation (Fig. 1, loc. XV), a number of fossils were collected by Dunbar, Keith, and the writer. From this locality, Cooper identifies the brachiopods *Syntrophina* sp., *Archaeorthis* sp., and *Clarkella* n.sp. The trilobites, identified by Raymond, are: *Cholopilus vermontanus* Raymond, *Petigurus cybele* (Billings), *Gignopeltis rara* (Billings), *Hystricurus mammatus* Raymond, *Pilekia eryx* (Billings), *Lloydia saffordi* (Billings), and *Leiostegium obtectum* Raymond. *Cholopilus vermontanus*, *Leiostegium obtectum*, and *Lloydia saffordi* are also found in the Highgate formation. The Grandge formation is probably hundreds of feet thick, and, for the present, Raymond regards it as of "early Beekmantown age".

Thirty feet above the base of the Grandge slate limestone blocks hold *Niobe* sp.

[101] P. E. Raymond: *Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont*, Geol. Soc. Am., Bull., vol. 48 (1937). In press.

DETAILED DESCRIPTION OF LOCALITIES

Highgate Falls, Missisquoi River.—At Highgate Falls, beneath the highway bridge (Fig. 1, loc. XI), massive dolomites occur in a low arch, which shows about 15 feet of thickness and makes the top of the falls and dam. Downstream, or west of the dam, there is more of the same kind of dolomite, but it is in a greatly disturbed condition and is thrust faulted over the Highgate slates below. The whole is a thrust series of thin- and thick-bedded gray dolomites, rare intraformational dolomite conglomerates, and thin zones of black slate. A part of the nearly horizontal thrust plane is clearly visible for several hundred feet along the north bank of the river. In a calcareous shale of this thrust block, Cecil H. Kindle collected in 1934 a number of brachiopods and more trilobites, which are said to be very much like those of the Gorge formation, proving that this thrust block repeats the Gorge formation below.

Downstream, and structurally beneath these Upper Cambrian dolomites, is the following descending sequence about 320 feet thick:

LOWEST ORDOVICIAN

UPPER HIGHGATE dark-blue banded slate, with a few thin beds of dolomite, and, toward the base, limestone and rare limestone conglomerate. Observed thickness, about 220 feet, with the top faulted off. The detail is:

Bed		Thickness Feet
16.	Dark-blue slate, dipping to the east	25
15.	Dark-blue, thin-bedded, brecciated dolomite	12
14.	Laminated dolomite, weathering yellowish	1
13.	Black slates and thin-bedded limestone	5
12.	Dark-blue banded slate, with torn and moved blocks of yellow-weathering dolomite in two beds, one 10 inches, the other 18 inches, thick. Slate much squeezed and crumpled. Changes gradually into lower Highgate division. Thickness cannot be measured, estimated at	175

LOWER HIGHGATE thin-bedded and banded limestones, dolomites, and limestone breccias. Thickness more than 100 feet. In detail as follows:

11.	Dark-blue, banded, impure limestone, much cleaved, dipping about 70 degrees to the southeast. At other localities the entire lower Highgate is of this nature. About	25
10.	Knobbly, calcareous, cleaved slates	3 – 4
9.	Calcareous, cleaved slate, with a few limestone layers, the slate replete with intraformational angular limestone pieces averaging less than 3 inches across	5 – 6
8.	Blue, thin-bedded limestone	1
7.	Yellowish-weathering, massive, banded dolomite	1 – 2
6.	Blue, thin-bedded limestone	1
5.	Yellowish-weathering, massive, banded dolomite	5
4.	Thin-bedded, banded limestone	15

The upper part of this zone on the south bank of the river and to the west of the power plant, includes two superposed erratic blocks of bioherm limestone, the upper 10 feet long and 4 feet thick, the lower 4 feet long by 4 feet thick. The thin-bedded limestone is squeezed over and around these erratics. How they got into their present position is unknown.

Bed		Thickness (Feet)
3.	Two massive, banded, conglomeratic (intraformational) dolomite beds, with thin-bedded limestone inclusions as much as a foot wide. Upper bed, 4 feet thick; lower, 2 feet. On the south bank these beds swell out to	12
2.	Thin-bedded and banded dolomitic strata. It should be noted here that on the north bank of the river there are two conglomeratic zones separated by the 10 feet of dolomite. The south bank has only one such zone, next to be described.	10
	BASAL HIGHGATE THRUST BRECCIA	26
1.	This thrust breccia dips under the power house on the south bank of the river. To the east of the plant, on the south bank, it looks like the regulation Mill River or the Rockledge limestone breccia, but it has an abundance of small pieces of blue, thin-bedded limestone derived out of the Gorge formation; it is about 10 feet thick, but within a few hundred yards to the west the beds swell out into a great limestone breccia, 26 feet thick. The base of the zone shows that it slid over the Gorge formation and picked up parts of it (Pl. 6, fig. 2). The great bed has large angular and subangular blocks of dolomite and limestone, some of which are 12 feet long. These large blocks make up at least half of the bed and appear to be broken-up dolomite strata and limestone lentils of the Highgate formation itself. The remainder of the material is dark-blue, thin-bedded limestone picked up from the upper Gorge beds. The whole is cemented by a dolomite paste full of rounded sand.	

The continuation of this section is given with the description of the section in Highgate Gorge, at the end of the Upper Cambrian discussion.

All these strata Keith correctly refers to the Highgate formation, but in 1932 he correlated the lower limestone breccia with the Mill River conglomerate. The latter, however, occurs everywhere at the base of the Hungerford, and not at the base of the Highgate and above the Gorge, as is the case here.

Highgate Center.—The lower Highgate is exposed just west of the village and to the north of an east-west road, in a pasture (Fig. 1, loc. XXVII), with fossils in the interbedded thin limestone and slate. Keith re-discovered this locality of Walcott's, and the latter pronounced the fossils collected by Keith to be of Upper Cambrian age. *Lingulella acuminata* is especially common. Raymond identifies from this locality *Leiostegium puteatum* and other trilobites.

The low railway cut a quarter of a mile farther to the north has a fine exposure of about 50 feet of the alternating thin-bedded shales and limestones of the lower part of the Highgate formation. All the beds are highly cleaved, making it impossible to get more than an occasional trilobite, and none of those found has been determinable.

This limestone exhibits its extraordinary banded nature clearly on weathered surfaces in the railway cut, and Keith, Sayles, and the writer counted, in a thickness of 35 feet, about 2,830 bands; there are about 81 bands in each foot of limestone, but some of the bands are half an inch across and even 1 or 2 inches thick. This place also exposes one dolomite bed 4.5 inches thick, which is not banded.

One and one-fourth miles northeast of Highgate Center.—A roadside cutting in the lower Highgate (Fig. 1, loc. XXXIV) yielded two small species of *Lingulella* and *Acrotreta*, and the trilobites *Lloydia seelyi* and *Leptopilus declivis*.

Two miles northeast of Highgate Center.—Northeast of Highgate Center (Fig. 1, loc. XXXV), the lower Highgate is exposed in a roadside cutting. Although fossils are common, they are very poor, and most of them are not determinable. They include trilobites, small gastropods, rare cephalopods, and many cystid plates and columnals; the brachiopods are represented by *Clarkella* sp. and *Orthis?* cf. *O. euryone*.

One-fourth of a mile farther north, along the roadside, a reef limestone lentil about 600 feet long and 30 feet or more thick is exposed, in which diagenetically altered trilobite fragments are common. Orthids and cystid plates are also to be found at this locality. In places, the lentil is a fine intraformational conglomerate.

Four miles north of Highgate, and just west of Rock River, the lower Highgate yields *Lingulella acuminata*.

Two and one-third miles northwest of Highgate Center.—Two and one-third miles N. 15° W of Highgate Center (Fig. 1, loc. XXXVI), a limestone ridge of the lower Highgate formation strikes northeast across an east-west country road, and here thin-bedded limestones are interbedded with calcareous shales, all folded in a syncline. An unnamed trilobite was collected in black slates. These slates are banded and show from 30 to 40 feet of thickness. Higher, interbedded thin limestone and slate appear, each bed varying in thickness from 0.5 to 6 inches. There are also lenses of intraformational conglomerates, some of which are 5 inches thick and from several feet up to many yards long. Their cement has more or less rounded sand grains. These thin limestones have an aggregate thickness of between 60 and 75 feet. Fossils are hard to collect because of the cleavage, but in places they are abundant and consist mainly of undetermined trilobites and brachiopods.

Dolomite of the lower Gorge division is exposed about 500 feet to the west of these lower Highgate limestones.

Southeast of Burlington.—According to Keith,[102] the type area of the Williston limestone is in "the western part of the township of Williston about 5 miles southeast of Burlington." Six miles southeast of Burlington, beside a schoolhouse on the Hinesburg road, the typical Williston formation may be seen; to the writer, it looks like the regulation massive and banded magnesian limestone of Beekmantown age found in the Western Sequence. Apparently all the formations of the Central Sequence are here eroded away. Some of the beds are crystalline or crinoidal limestone, with here and there a dark-blue dolomitic zone from one to several feet thick. In one of the crinoidal limestones, Dunbar and the writer and, later, Raymond collected the trilobite *Lloydia saffordi* and a pygidium like that of *Platycolpus*. The brachiopod *Archaeorthis* cf. *A. electra* also occurs here. These species show clearly that the age of this Williston limestone is Beekmantown, and it is apparently well up in the Canadian series.

[102] Arthur Keith: *Stratigraphy and structure of northwestern Vermont*, Washington Acad. Sci., Jour., vol. 22 (1932) p. 377.

Peabody Museum, Yale University, New Haven, Conn.
Manuscript received by the Secretary of the Society, November 29, 1935.
Presented before the Society, December 29, 1934.